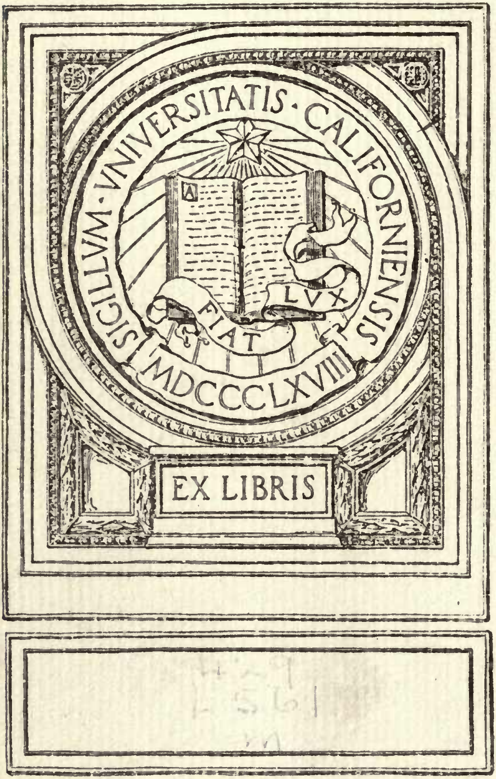
METEOROLOGY

R. G. K. LEMPFERT



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METEOROLOGY

METEOROLOGY

BY

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WITH 48 ILLUSTRATIONS AND DIAGRAMS

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PREFACE

DURING the twenty-seven years which have elapsed since the publication of Dr. H. N. Dickson's volume on "Meteorology" in the University Extension Series, the organised exploration of the atmosphere by means of kites and balloons, which began just about the time when the book was published, has placed at the disposal of meteorologists much observational material of an entirely new kind. The progress made on the theoretical side has been correspondingly rapid. In the following pages it has been my endeavour to present in simple form the main results which have been derived from the new method of observation and to fit them into their proper place in the scheme of dynamical meteorology. The observational side of the subject has been omitted. For descriptions of meteorological instruments and methods of using them, the reader should consult books of instruction in observing, such as the "Observer's Handbook," prepared by the Meteorological Office. The climatological aspect has also been omitted.

The reader may consider that some apology is due for the confusion of units used. In the present state of the subject any attempt to use only one system throughout would appear forced. I have therefore, as a rule, given the examples in the units in which I happened to find the data. Thus it comes about that temperatures are expressed sometimes on the Fahrenheit scale and sometimes in absolute or Centigrade

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degrees, while wind velocities are given in miles per hour or metres per second. Pressures have generally been given in millibars, but in examples taken from maps which date from before the official adoption of the millibar, the specification in inches of mercury has been allowed to stand. A table of equivalents in millibars of mercury inches is added at the end of the volume.

I have to express my thanks to the Director of the Meteorological Office for permission to use published and unpublished material available at the Office, permission of which I have availed myself freely. I have also to thank the Controller of His Majesty's Stationery Office for permission to reproduce the following illustrations from official publications of the Meteorological Office: From the Weather Map, M.O. 225 i, Figs. 3, 4, 5, 6; Glossary, M.O. 225 ii, Figs. 39, 41, 42, 48; Cloud-Forms, M.O. 233, Figs. 13, 14, 15, 16, 17, 18, 19, 20; the Estimation of Height from Readings of an Altimeter, M.O. 228, Fig. 11.

I am indebted to the Council of the Royal Meteorological Society for permission to reproduce Figs. 31 to 38 from the Society's Quarterly Journal, and for the loan of the blocks, and hereby express my thanks for the courtesy. In conclusion, I desire to thank my colleague, Mr. J. S. Dines, M.A., for looking through the proof-sheets and for several valuable suggestions and criticisms; and Messrs. A. T. Bench and A. G. W. Howard, of the Meteorological Office, for assistance in preparing the illustrations.

R. G. K. L.

July, 1920.

CONTENTS

CHAPTER I

INTRODUCTORY: WEATHER MAPS

PAGES
Buys Ballot's Law—Anticyclones—Cyclones—V-shaped Depressions—Secondaries—Wedges - - - - 1—23

CHAPTER II

PRESSURE

The Barometer—Measurement of Pressure in Millibars—Reduction to Mean Sea Level—Determination of Height from Pressure - - - - - - - 24—35

CHAPTER III

WIND

Beaufort Scale of Wind Force—Measurement by Anemometer—Gustiness and Exposure—Measurement of Wind at High Altitudes by Pilot Balloons and other Methods - - 36—44

CHAPTER IV

TEMPERATURE

Exposure of Thermometers—Measurements of Temperature in the Upper Air—Temperature Scales—The Absolute Scale 45—50

CHAPTER V

CLOUDS

The International Classification of Cloud Forms—Determination of the Height of Clouds - - - - - 51—57

CHAPTER VI

THE RELATION OF WIND TO PRESSURE DISTRIBUTION

The Gradient Wind—Cyclostrophic and Geostrophic Components—Steady Flow of Air over Long Distances—The General Circulation of Air over the Globe at Sea Level and at 4,000 metres 58—75

CHAPTER VII

THE TRAJECTORIES OF SURFACE AIR CURRENTS

PAGES
Trajectories in Cyclonic Depressions—Their Connection with Rainfall - - - - - - - - 76—89

CHAPTER VIII

LINE SQUALL PHENOMENA

Isochronous Lines—The Interference of Air Currents—Lines of Flow—Lines of Convergence and Divergence - - 90—100

CHAPTER IX

DIURNAL VARIATION OF WIND AND THE VARIATION OF WIND WITH HEIGHT ABOVE THE GROUND

Eddy Motion—The Gradient Wind as a Measure of the Air Flow at 1,500 feet - - - - - - 101—110

CHAPTER X

THE WINDS OF THE FREE ATMOSPHERE

Computation of the Pressure Gradient at High Levels - 111—120

CHAPTER XI ✓

THE PHYSICAL PROCESSES OF THE ATMOSPHERE

Changes associated with Change of Level—Lapse Rate—Formation of Cloud and Rain—Rate of Fall of Raindrops - 121—129

CHAPTER XII ✓

THE VARIATION OF TEMPERATURE WITH HEIGHT IN THE TROPOSPHERE

Inversions—Formation of Fog in Valleys—Sea Fog—Effect of Eddy Motion on the Lapse of Temperature—Inversions over Cloud - - - - - - - 130—140

CHAPTER XIII ✓

THE STRATOSPHERE

Variations in Level of the Boundary between Stratosphere and Troposphere—Variations of Temperature in the Stratosphere—Origin of the Stratosphere - - - - 141—149

CHAPTER XIV

THE THERMAL STRUCTURE OF THE ATMOSPHERE

PAGES
The Seasonal Variation of Temperature at Different Levels—The Lapse of Temperature in Cyclones and Anticyclones—Winds in the Stratosphere - - - - - 150—161

CHAPTER XV

THE ORIGIN OF CHANGES OF PRESSURE

Correlation between Pressure and Temperature in the Troposphere—Correlation between Pressure at 9 Kilometres and Temperature in the Troposphere - - - - - 162—167

CHAPTER XVI 4

TROPICAL REVOLVING STORMS

Their Seasonal Variation and other Characteristics—Tornadoes 168—176

CHAPTER XVII ✓

THE CHEMICAL CONSTITUTION OF THE ATMOSPHERE

Changes in Composition at Great Heights - - - 177—181
Pressure, Table of Equivalents - - - - 182—183

INDEX - - - - - - - 184—186

LIST OF PLATES

FIG. PAGE

3. DISTRIBUTION OF RAIN AND CLOUD IN CYCLONE, NOVEMBER 12, 1901 - - - - *facing* 10
4. DISTRIBUTION OF RAIN AND CLOUD IN CYCLONE, MARCH 24, 1902 - - - - - ,, 10
5. DISTRIBUTION OF RAIN AND CLOUD IN CYCLONE, SEPTEMBER 10, 1903 - - - - - ,, 10
6. DISTRIBUTION OF RAIN AND CLOUD IN CYCLONE, NOVEMBER 12, 1915 - - - - - ,, 10
12. PYRTON HILL WIND RECORD, DECEMBER 11–12, 1912 ,, 38
13. CIRRUS AND RIPPLED CIRRO-CUMULUS CLOUD - - *following* 56
14. RIPPLED CIRRO-CUMULUS CLOUD - - - ,, 56
15. CIRRO-CUMULUS CLOUD BECOMING ALTO-CUMULUS - ,, 56
16. ALTO-CUMULUS CLOUD BECOMING STRATO-CUMULUS - ,, 56
17. STRATO-CUMULUS CLOUD FROM BELOW - - ,, 56
18. STRATO-CUMULUS CLOUD FROM AN AEROPLANE (4,000 FEET) - - - - - - ,, 56
19. CUMULO-NIMBUS CLOUD - - - - ,, 56
20. VALLEY FILLED WITH FOG, 300 FEET DEEP - - ,, 56
39. LINE-SQUALL AT ABERDEEN, OCTOBER 14, 1912 - *facing* 97

METEOROLOGY

CHAPTER I

WEATHER MAPS

DURING the Great War the military authorities, realising the importance of withholding from the enemy all knowledge of the weather conditions in the area over which the Allies had control, insisted on the suppression of all weather reports in the newspapers, but almost immediately upon the termination of hostilities the *Morning Post* started the regular publication of a weather map and shortly after *The Times* resumed publication on an enlarged scale of the maps which had appeared in its columns for many years. The *Daily Telegraph* followed. This action on the part of our leading daily newspapers testified to the interest which they expected their readers to take in the ordered study of the weather. They argued rightly that such interest had been much quickened by the obvious influence of weather conditions on aviation and all that that had meant to the general public during the years of struggle, but aviation is by no means the only part of military operations in which the weather may have a deciding influence. Military history from the earliest times abounds with instances of weather conditions profoundly modifying a campaign, generally as much by their effect on transport as by their direct action on the troops.

In the late campaign the use of poison gas opened up a new field for the application of meteorology, since its use was only practicable under very special conditions of wind. Moreover the atmospheric conditions prevailing up to great heights affect materially the flight of modern long range projectiles and the meteorologist, who nowadays does not confine himself to observing and studying the conditions near the surface was able to supply much valuable information to the gunners. Weather study has thus come to occupy a much more prominent position as a result of the War.

The Weather Map is the foundation on which much of the science of meteorology rests. Modern forecasting is entirely dependent on maps, but it is not merely from the point of view of forecasting that the map makes its appeal. To him who has learned to read it it gives a succinct summary of the weather conditions over a large area and as such is of practical value to everyone whose business may be affected by weather conditions on a wide scale.

Let us examine these maps more closely and see how they are put together. They are based on observations made at a number of stations and collected together at a central office. Each station reports among other things, the height of the barometer, the direction and strength of the wind, the temperature of the air and the state of the weather, whether it be bright, cloudy, rainy, foggy and so forth. It is essential that the information should be collected as rapidly as telegraph, telephone or wireless can bring it together. In north-western Europe 7 a.m. of Greenwich mean time has been adopted as the hour of morning observations. Punctually at that hour reports are prepared and dispatched to the Meteorological Office from over 40 stations in the British Isles. By 8 a.m. the British information is generally complete and under favourable circumstances it may be supplemented by wireless reports from ships on the Atlantic. Immediately upon receipt of the home reports a so-called collective message giving a selection of the British reports is prepared for transmission to foreign meteorological services in exchange for similar reports from abroad. Shortly after, the foreign reports begin to arrive but several hours necessarily elapse before the information from these sources is complete.

The whole set of reports received is set out *in extenso* in the *Daily Weather Report* of the Meteorological Office which is issued later in the day. The preparation of a weather map thus involves close co-operation between many hundreds of persons, observers, telegraphists, telephone clerks and wireless operators as well as the staff at the central Office and it is the constant endeavour of those responsible for the organisation to speed up the interchange of information and extend its scope.

Immediately upon the receipt of each telegram the information which it contains is plotted on a large scale outline map of Europe on which the positions of the stations have been marked in advance. Against each station is written the appropriate barometer reading, and temperature, the wind is represented by an arrow, its strength being indicated according to the Beaufort Scale of wind force (see p. 37) by the number of flèches on the end of the arrow, or if we prefer it by specifying its velocity in miles per hour. The weather is indicated by letters or by conventional symbols. The weather notation most frequently adopted is that suggested by Admiral Beaufort in 1805, to which a few additions have been made more recently.

b = blue sky
bc = blue sky with some cloud
c = sky cloudy
o = sky overcast
d = drizzle
e = wet air
f = fog
g = gloomy appearance
h = hail
l = lightning
m = mist

p = passing showers
q = squalls
r = rain
s = snow
t = thunder
u = ugly appearance of sky
v = unusually good visibility
w = dew
x = hoar frost
y = dry air
z = haze

A system of symbols, adopted internationally by agreement between the meteorological services of different countries may be used alternatively or in conjunction with the Beaufort letters:—

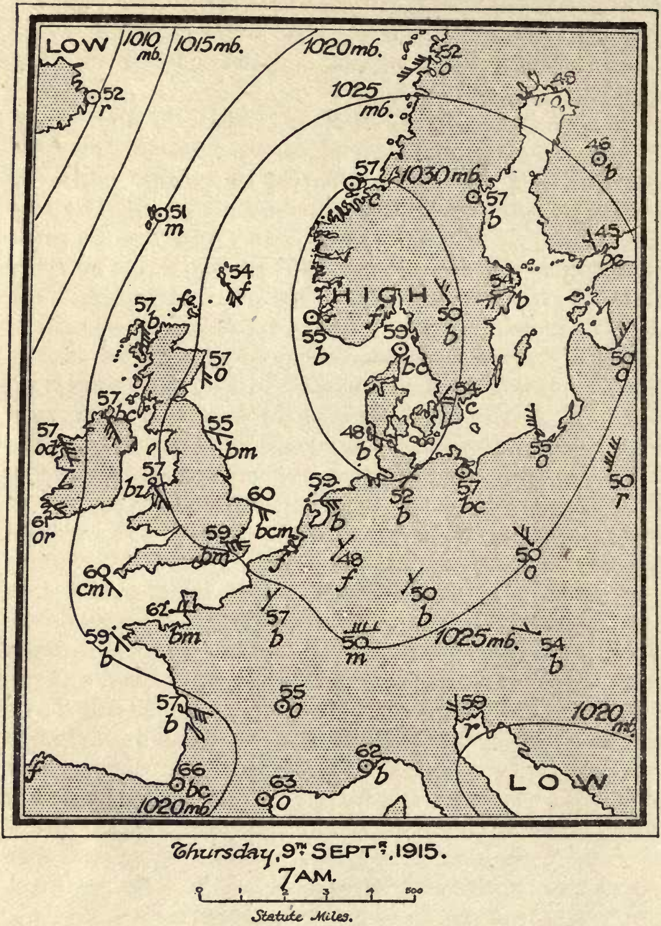
● = rain
✳ = snow
▲ = hail
△ = soft hail

/ = gale
↯ = lightning
T = thunder
TR = thunderstorm
≡ = fog

The completed map, or synoptic chart as it is sometimes called thus gives a bird's eye view of the weather conditions over a wide area from which conclusions may be drawn as to past or future conditions. Such maps are now drawn at the Meteorological Office four times a day, at 1 a.m., 7 a.m., 1 p.m., and 6 p.m.

Despite the fact that maps have now been drawn day by day for over half a century we may safely say that no two maps have been identical. Nevertheless on any individual map the observations show a certain orderliness, they do not arrange themselves haphazard. If we fix our attention on the observations of weather we can generally find considerable areas of fine weather or of overcast sky or again well-marked areas over which rain is falling. The winds also generally show wide sweeping currents affecting large areas, but it is in the barometer readings that the greatest orderliness is found. So great is this orderliness that individual barometer readings are not as a rule entered on the maps published in the newspapers. Instead the barometric situation is represented by a series of lines like the contour lines which show heights on the maps of the Ordnance Survey. The lines are so drawn that the barometer would show the same reading at any point along a given line and are called *isobars*. The configuration of the isobars shows at a glance the regions where the barometer is high and those where it is low and in fact the configuration of the isobars forms the basis of the classification of weather maps which is the foundation of forecasting, and has provided the groundwork of a great part of modern meteorology.

No close scrutiny of weather maps is needed to disclose the fact that there is an intimate relation between the picture presented by the isobars and the distribution of wind and weather. As regards the former almost any map will reveal a connexion between the strength of the wind and the spacing of the isobars. Where the isobars are close together there the wind tends to be strong and *vice versa*. As regards wind direction we need not look far for examples of the generalisation known as Buys Ballot's law "Stand with your back to the wind, the region of lowest barometer will then be on your left but slightly in front of you." In the Southern hemisphere the relation is reversed, the lowest barometer is on the observer's right, while near the equator the law does not apply. We proceed to examine some of the more typical configurations of isobars. Fig. 1 shows the map for the morning of September 9th, 1915, and may be taken as a typical example of the distribution to which the name *anticyclone* is applied. The isobars form closed curves with the highest barometer readings at the centre of the system. The central isobar, which encloses Denmark, Southern Norway and a part of Sweden has the value 1030 millibars, the equivalent of 30·42 inches; the highest barometer reading reported was 1031·1 millibars at Christiania. As is usual in anticyclones the isobars are rather widely spaced, the line of 1025 millibars, the next highest value shown on the map, embraces all the central part of the map. Theory indicates that the centre of an anticyclone should be a region of no air motion and the actual observations show calms, indicated symbolically by circles, at most stations in Denmark and Southern



Thursday, 9th Septr., 1915.
7 AM.

Fig. 1

Norway. In other parts of the map we have no difficulty in tracing a circulation round the centre in a clockwise sense as we should expect from Buys

Ballot's law to which reference has just been made. Over the Baltic area the wind is from North, over Germany and France it becomes easterly, turns into south-east and south over Scotland and Ireland and gets round to south-west or even west in the North of Norway and Sweden. In force the winds are generally light, a normal feature of anticyclones, as the isobars are generally widely spaced. The majority of the wind arrows have only one or two or at most three flèches. Only near the edge of the map where the isobars are more closely spaced do we find stronger winds, four or five flèches at Malin Head and in the Hebrides, six at Bodo on the coast of Norway near the boundary of the map and five at Warsaw on its eastern margin. Of the weather in an anticyclone we may say that rain is generally absent. Fine weather with very little cloud is characteristic of many anticyclones but a type of anticyclone in which the sky is covered with low structureless cloud must also be recognised. Sometimes this cloud is very heavy and gives rise to a gloominess which suggests the imminence of rain. Our example is of the cloud free type. Most stations reported the weather as *b* or *bc*, blue sky or blue sky with some detached cloud. The only observations of rain are from the extreme margins of the map, on the west coast of Ireland and at Warsaw, regions which may be regarded as outside the anticyclonic area. We note, however, a number of observations of fog, a phenomenon which often occurs in anticyclonic weather. The symbol for fog appears on our map in Orkney and Shetland and also at Flushing, Christiania, Corunna and at some German stations.

The temperature under anticyclonic conditions depends to a great extent on the cloudiness. In the cloudless type in summer very high temperature may be experienced as a result of the unchecked absorption of the sun's heat during the long summer day, while in winter, when night predominates absence of cloud allows of loss of heat by radiation and intense cold may result. A heavy cloud layer checks both the inward and outward flow of heat, and under such conditions temperature is generally not extreme in either direction. Cold weather may, however, be experienced under cloudy anticyclonic conditions in winter when an anticyclone is centred over the North Sea and the consequent east wind brings very cold air to our shores from the storehouse of winter cold of the Eurasian Continent. Anticyclones belong to our most stable types of weather. They often persist for days or sometimes even weeks without any great change taking place in the general conditions. Sometimes anticyclones exhibit the phenomenon of travel which we shall have occasion to discuss more fully in connexion with other weather types. The whole system moves slowly across the map carrying with it its characteristic distribution of winds and weather.

The anticyclone of September 1915 showed permanence combined with slow travel. The anticyclonic type persisted from September 5th until the 14th. At the beginning of that period, the region of highest barometer was over France. As successive maps were worked out it was found to be gradually moving north-eastward, until on the 9th, the day of our map, it was over the Skagerak; subsequently the centre moved slowly eastward and by the 12th of the month passed away into Russia and the system gradually disintegrated.

Fig. 2 shows an example of a totally different type to which the name *cyclone*, *depression*, or even, for brevity "*low*" has been given. The map is one of special interest for it represents the meteorological distribution on the morning of February 17th, 1915, on the evening of which day two German airships were lost through stress of weather on the coast of Denmark.

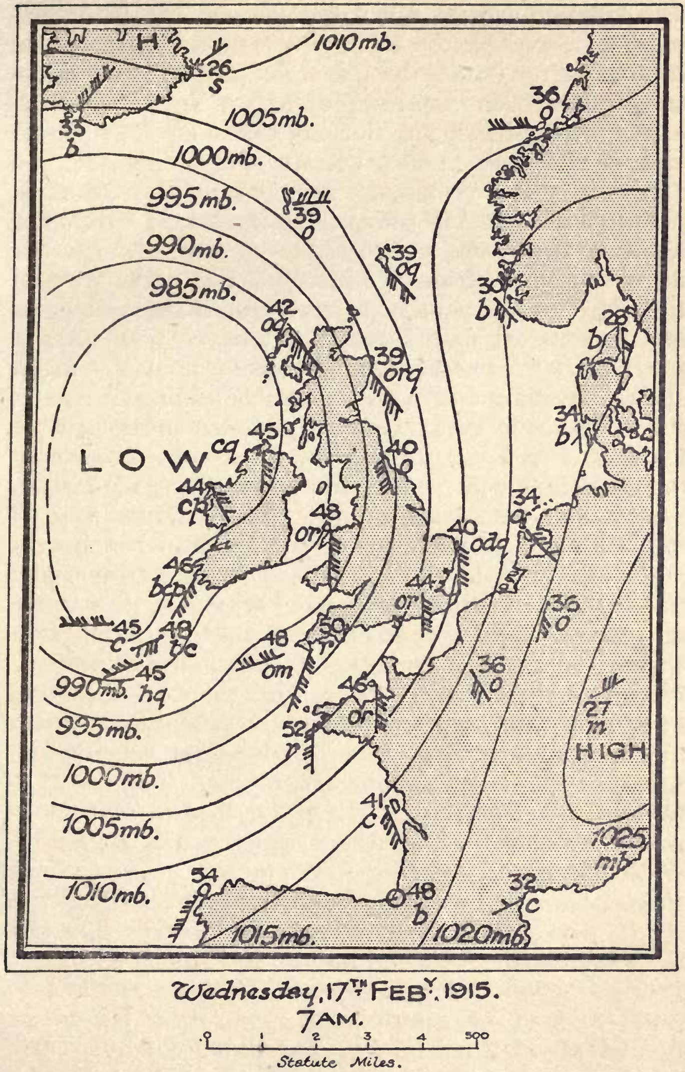

1010mb.
1005mb.
1000mb.
995mb.
990mb.
985mb.
LOW
HIGH
1015mb.
1020mb.
1025 mb.
26 s
35 b
36 o
39 o
39 oq
42 oq
39 orq
30 b
28 b
34 b
45 cq
44 cp
40 o
48 or
40 odq
34 or
46 bcp
44 or
45 c
48 bc
36 o
45 hq
48 om
50 r
36 o
27 m
46 or
52 r
41 c
54 o
48 b
32 c
Wednesday, 17th Feby. 1915.
7AM.
0 1 2 3 4 500
Statute Miles.

Fig. 2

The occasion exemplifies, if that be needed, the importance of guarding the secrecy of weather information in warfare for it may be taken for granted that if the enemy had been cognisant of only a fraction of the information shown on our map the airships would have been recalled before disaster overtook them.

As in the anticyclone, the isobars form closed curves, but we notice that in the cyclone the lowest barometer readings are at the centre of the system. The isobar for 985 mb. encloses a region off the north-west of Ireland. In shape the isobars are rather irregular and in contrast with the anticyclone they are closely spaced. The winds circle round the centre in a counter-clockwise sense and offer an excellent illustration of Buys Ballot's law. They are from north-east in Iceland, from east in Faröe, from south-east over Scotland and south-south-east or south over France, England and the Netherlands. In the South-west of England and in Ireland they draw into south-west and over the Atlantic, 200 miles from the Irish coast, we find an observation of a wind from due west. We also find general confirmation of the suggestion that where the isobars are close together the winds are strong and *vice versa*. They are light on the eastern margin of the map where the isobars are relatively far apart. Over the British Isles they are strong, reaching gale force at several stations. The distribution of weather is interesting and typical of that in many cyclones. In Southern Norway and Denmark, which may be regarded as on the margin of the cyclone the weather is fine and the sky almost free from cloud. Further westward, there is increasing cloud. The sky is overcast over Shetland and at the Dutch and most French stations. Proceeding still further westward we find an area of steady rain; rain was falling at almost all stations in England and Scotland. Still further west—in Ireland, we find very "mixed"

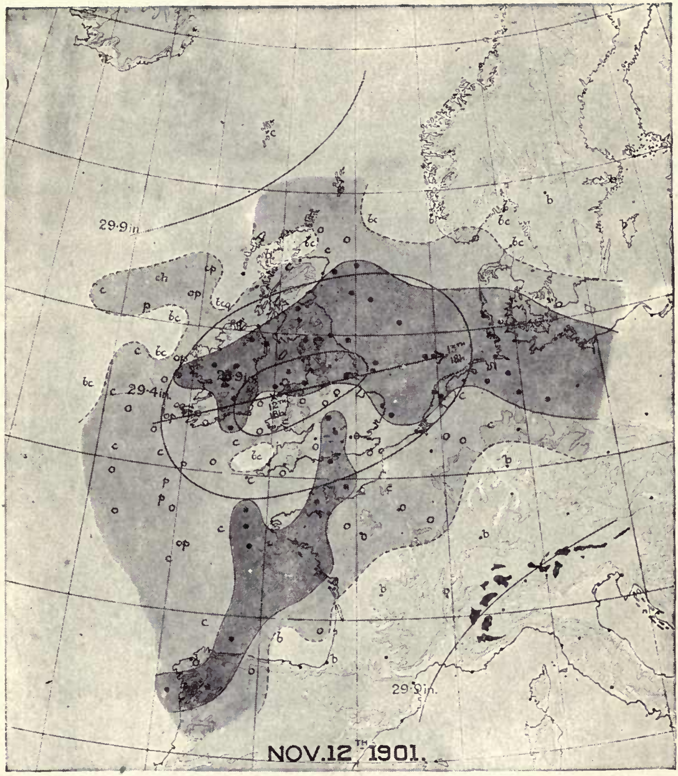


FIG. 3—DISTRIBUTION OF RAIN AND CLOUD IN CYCLONE

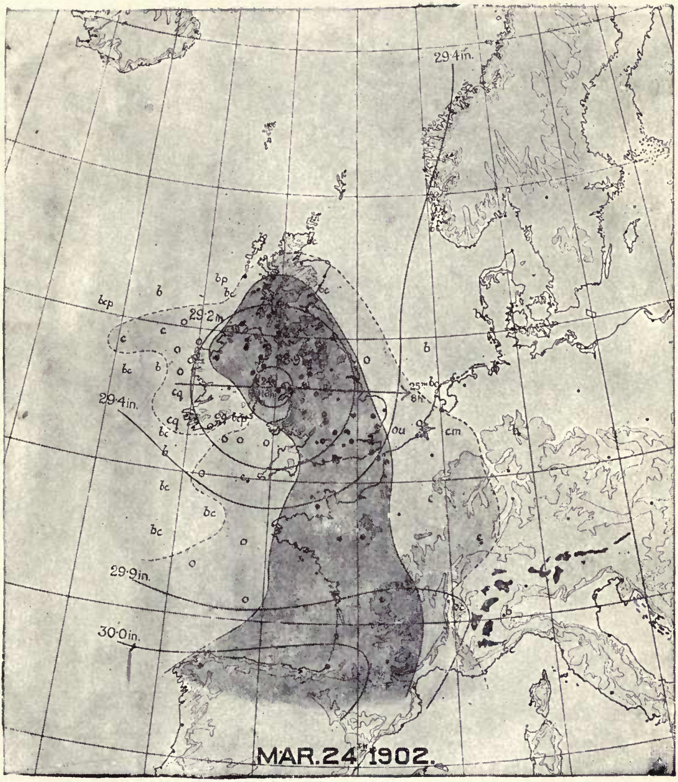
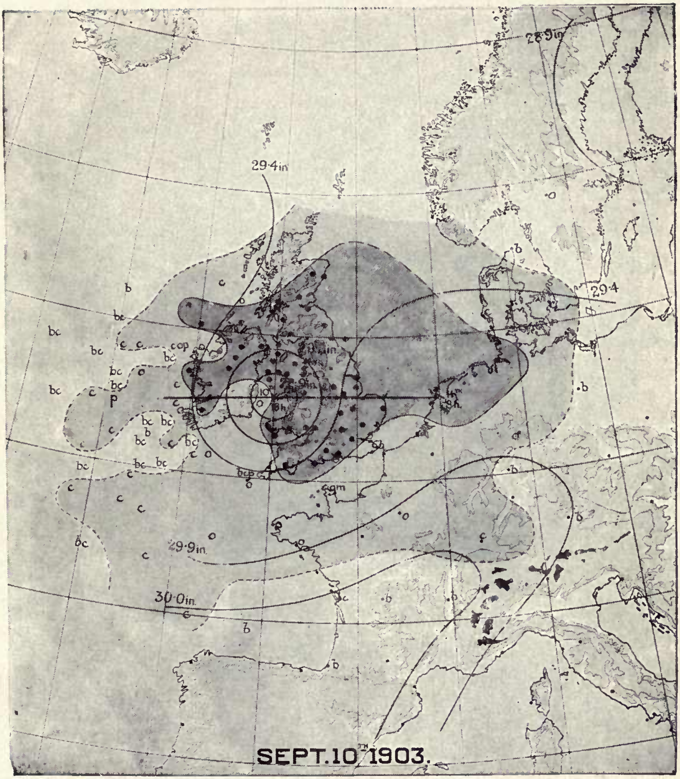


FIG. 4—DISTRIBUTION OF RAIN AND CLOUD IN CYCLONE



SEPT. 10TH 1903.

FIG. 5—DISTRIBUTION OF RAIN AND CLOUD IN CYCLONE

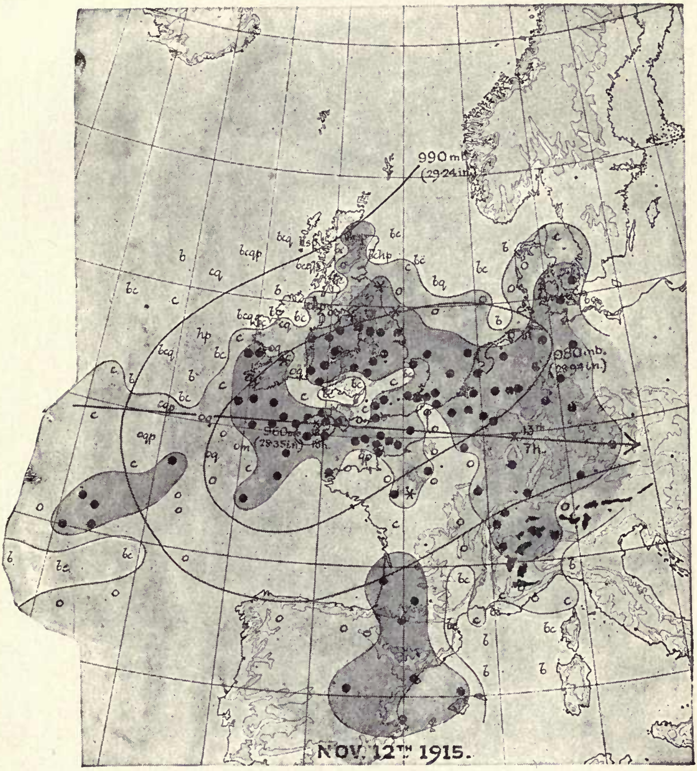


Fig. 6—Distribution of Rain and Cloud in Cyclone

weather, the sky is mostly overcast, but at Valencia in the extreme south-west it has cleared partially. Showers are reported from various places. Out in the Atlantic, a ship reports hail showers, a neighbouring one fair weather.

The distribution of wind and weather shown on the map of Fig. 2 may be regarded as typical of most cyclones, though the details will vary on different occasions, for example the area of steady rain may be of greater or less extent and may vary somewhat in position with regard to the point of lowest barometer, as will be evident from other cases which we shall consider. Examples are given in Figs. 3, 4, 5 and 6 of generalised maps which show the distribution of cloud and rain with regard to the isobars for various typical cases. The dark shading represents the area within which continuous rain was falling, the lighter shading the region in which the sky was covered by cloud. The Beaufort letter *p* shows the positions of isolated showers. The weather shading, indicating cloud and rain, must be distinguished from the contour shading over the land, for heights of 200 and 3,000 metres. Though the details differ considerably we recognise on all the maps an area of cloud and rain extending further from the centre of the system of isobars on the eastern than on the western side, and a region of clearing sky and showers on the western side of the centre.

Let us now return to the map for February 17th, 1915 (Fig. 2) and examine its subsequent developments. If the reader will imagine the whole system displaced bodily towards the north-east, so that the centre of the innermost isobar, instead of being just off the north-west of Ireland is somewhere near the Hebrides he will get an idea of the state of affairs at 6 p.m., 11 hours after the time for which our map is drawn. If he imagine also that the barometer near the centre of the system has become somewhat lower, 980 millibars instead of 985, and that the whole system has grown somewhat larger, he will get a still better idea of what actually happened. The most obvious result of such a change would be that strong south or south-south-east winds and gales would have spread to Denmark and the eastern part of the North Sea. The weather there would change from fine to rainy or, as actually happened in this case, to snow. At the same time the "mixed" showery weather which our map shows over Ireland would be carried forward to England accompanied by a shift of wind towards south-west. That is precisely what happened, and we see here general principles on which forecasts may be based. Given one map, the forecaster has to estimate the displacement of the various systems shown thereon and the probable changes in their structure, and the rest follows. Unfortunately both may falsify anticipations. Let us put ourselves for a moment in the position of the German forecasters, for their position was very similar to that in which a British forecaster often finds himself unless he happens to have adequate wireless reports from the Atlantic. Over Germany the conditions were anticyclonic, high barometer, light winds and fine weather. On the western margin the Dutch stations, the most westerly from which reports were accessible, showed overcast skies, southerly wind and a falling barometer, precisely the same sort of reports which we so often get from our stations on the west coast of Ireland. Clearly a depression was indicated to the westward, but no information was available for estimating its character. Was it likely to prove deep, *i.e.* were the isobars closely packed and the winds correspondingly strong? Was it likely to be associated with much rain or snow? Would it move rapidly or slowly? To all these questions the German forecast officials could only give intelligent guesses, based on past experience, by way of answer. According to newspaper reports the German airships were observed by the Danish Lightkeepers at Blaavands Huk, a station on the coast of Jutland just north of the frontier between Slesvig and Denmark, at about 8 a.m. on February 17th, one hour after the time to which our map applies. We may imagine the German official forecast for their course to have read somewhat as follows: "Freshening southerly wind, fair early, some rain later," which may or may not have decided the military authorities to recall the airships by wireless. Very possibly they considered it safe to await developments as shown by the next map, and hoped that in the meantime the ships might accomplish their object, whatever that may have been. Had British reports been available a much more definite forecast such as "southerly gales, heavy rain or snow" would no doubt have been issued, and would have led to the instant recall of the ships; it might indeed have been issued twelve hours earlier, before the ships left their sheds.

We have considered the depression of February 17th, 1915, at some length because of its topical interest. In some respects the example is rather a complex one, as is indicated by the irregular shape of the isobars. A rather more simple case is illustrated in Fig. 27, which will be considered in greater detail later. In the cyclone of March 24th, 1902, there illustrated, the isobars are very approximately circles and the path of the centre of the system passed across the British Isles from west to east from Galway to the Wash, thus bringing all parts of the system within a part of the map for which full reports are available. The counter-clockwise circulation of the winds is complete. We may pause here to define some of the terms used in connexion with circular depressions. The track followed by the centre is generally referred to as the *path* of the depression. A line through the centre at right angles to the path is referred to as the *trough*. If the reader will think out the barometric changes which must take place at an individual station as the depression passes he will realise that the barometer will fall until the trough passes and will then commence to rise. To an observer at an isolated station the passage of the trough becomes evident by the cessation of the fall—at any rate if the depression is one of simple structure. Again if the depression is of simple structure the change from steady rain to the condition of "mixed" showery, but gradually improving weather often synchronises with the passage of the trough. The trough is also regarded as separating the "front" from the "rear" of the depresssion. It will also be obvious that the changes of wind experienced as a depression passes will be different according as the place of observation is on the right or left of the path. On the right hand (south) side of the path, the wind shifts from south-east, through south, and south-west to west or north-west as the depression passes. On the left hand (north) side of the path, the change is in the opposite direction from south-east through east, north-east, and north to north-west. The former change which in the Northern hemisphere is in the same direction as the motion of the sun is spoken of as *veering*, the latter as *backing*. The most frequent wind direction in this country is west or south-west, thus one of the earliest indications of the approach of a depression is often a *backing* of the wind from west or south-west towards south-east. Often this is observed before the barometer begins to fall.

The distribution of temperature in cyclones is rather irregular, being governed to a large extent by the special circumstances. The southerly and south-westerly winds are almost always, though not invariably warmer than the westerly and north-westerly winds in the rear of the system. In Fig. 2, the south-westerly winds over the Bay of Biscay in which temperature is between 50° and 55° are the warmest on the map, but the south winds on the eastern margin of the map, which draw their air supply from the cold land area of Central Europe are cold.

A depression may remain almost stationary for a considerable time, especially when it shows little intensity, that is to say when the isobars are not closely packed, but that is the exception rather than the rule. Normally they exhibit the phenomenon of travel. The path of the centre in the example shown in Fig. 26 is indicated by a series of crosses which mark the position at intervals of two hours. The rate of progression of the centre works out at 26 miles per hour. One of the main principles of forecasting is at once apparent from this idea of travel. If we may assume that the distribution within the cyclone remains constant a place situated to the south of the path, London for example, would experience in succession the conditions shown on a line drawn through it, parallel to the path. The barometer would fall, until the "trough" passed and would then commence to rise again. The record which a self-recording barograph would show, could in fact be reconstructed from the isobars of the map. The weather sequence would be increasing cloud, followed by steady rain, and after the passage of the trough, clearing weather with showers. The winds would first *back* towards south-east and subsequently *veer* to west and eventually to north-west. A place on the northern side of the path, would experience a somewhat similar sequence of weather but the winds would *back* from south-east through east, north-east and north to north-west. Unfortunately for the forecaster cases of such great simplicity are rare. Even if the path and rate of progress of the centre are correctly estimated, the internal configuration may alter. The depression may grow more deep or less deep, that is to say the barometric value at the centre may become lower or less low and the spacing of isobars may alter and the wind force vary in consequence. A depression which appeared likely to cause only fresh winds may develop and cause severe gales or *vice versa*, or again the area of rainfall or cloud may extend or contract. The amount of rainfall which a depression will bring is exceedingly difficult to estimate in advance. Sometimes the depression may split into two or develop a small subsidiary depression or secondary near its margin. The clear cut cases which lend themselves easily to generalisation are unfortunately rare.

The tracks or paths followed by depressions have formed the subject of much study. For many years past they have been worked out month by month and published in the *Monthly Weather Report* of the Meteorological Office. They form a most bewildering variety. The majority maintain their direction more or less throughout their course but cases of a sudden change are by no means uncommon and most baffling to the forecaster. The vast majority in the region of the British Isles show motion from west towards some easterly point, and the same is true in temperate latitudes the world over, but motion from east to west is by no means unknown. Almost any path seems to be a possible one; Fig. 7, reproduced from a Meteorological Office publication "The Weather of the British Coasts" gives some of the favourite tracks followed by depressions passing near the British Isles.

In Fig. 8 we have an example of another type, the so-called V-shaped depression. As its name implies, the isobars have the shape of the letter V. The central line, through the apex of the V's, is called the trough. The motion of the system is generally towards the east, the line of the trough remaining parallel to itself. In front of the trough, that is on its eastern side the wind

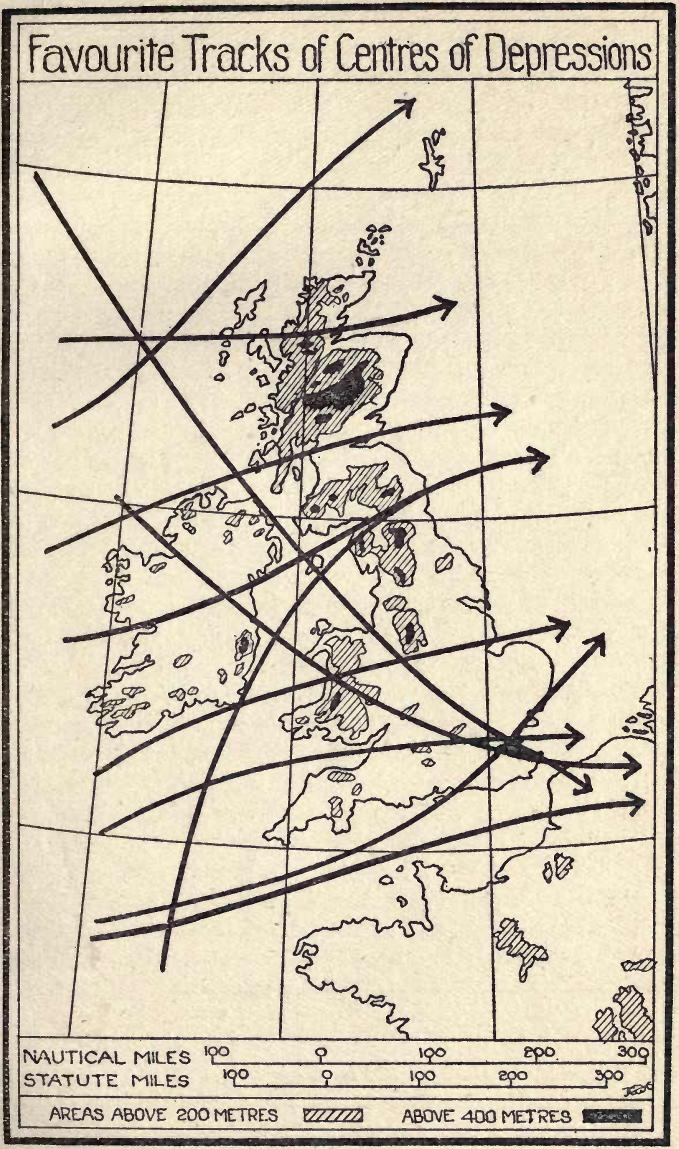
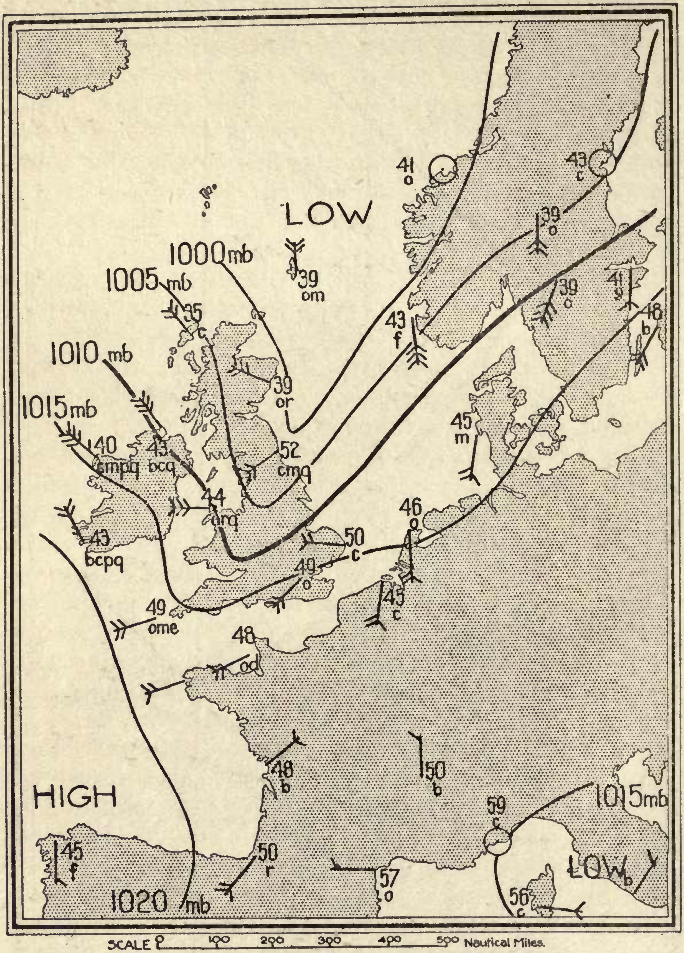

Favourite Tracks of Centres of Depressions
NAUTICAL MILES 100 0 100 200 300
STATUTE MILES 100 0 100 200 300
AREAS ABOVE 200 METRES ▨
ABOVE 400 METRES ■

FIG. 7

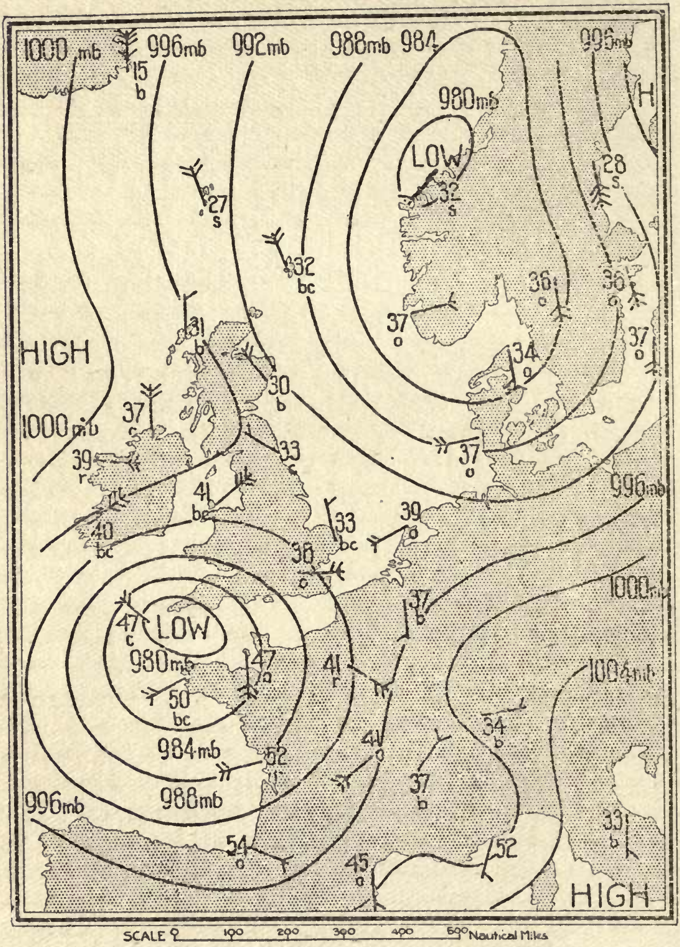


"V" Shaped Depression 7 a.m. Tuesday, 20th April, 1915.

FIG. 8

is from southerly directions, while in its rear it is from west, north-west or north. The sequence of weather experienced as a V-shaped depression passes by is similar to that in a circular one, increasing cloud, followed by rain until the trough is reached with a change to broken sky and showery weather after its passage. The change of wind experienced as the trough passes is often very sudden and accompanied by a marked increase in the rate of rainfall—the so-called clearing shower.

Mention was made when describing the internal changes which occur in depressions of the tendency to develop so-called "secondaries." If we like to so regard it, the V-shaped depression in Fig. 8 may be regarded as a "secondary" to a primary depression centred somewhere off the coast of Norway. Fig. 9 shows a different example of a secondary system. Here we have two well-marked centres and we might regard either as "secondary" to the other, though the term is often reserved for the more southerly partner, generally the less developed one, in such a binary system. Over Central England, between the two systems the winds are very light, but on the southern margin of the southern partner, where the isobars are closely packed they became very strong. The motion of such secondaries is generally in an easterly direction. In this particular case the system shown over the mouth of the English Channel moved south-eastward into the Mediterranean, where it gave rise to severe gales. The term "secondary" is one of the most loosely used in the forecasters' vocabulary. It is applied to easily identifiable systems such as those which we have considered in Figs. 8 and 9, and equally to systems which appear on our maps merely as slight distortions of the isobars which are apt to be entirely overlooked unless the isobars are drawn to fit the barometer readings with great accuracy. Such small distortions



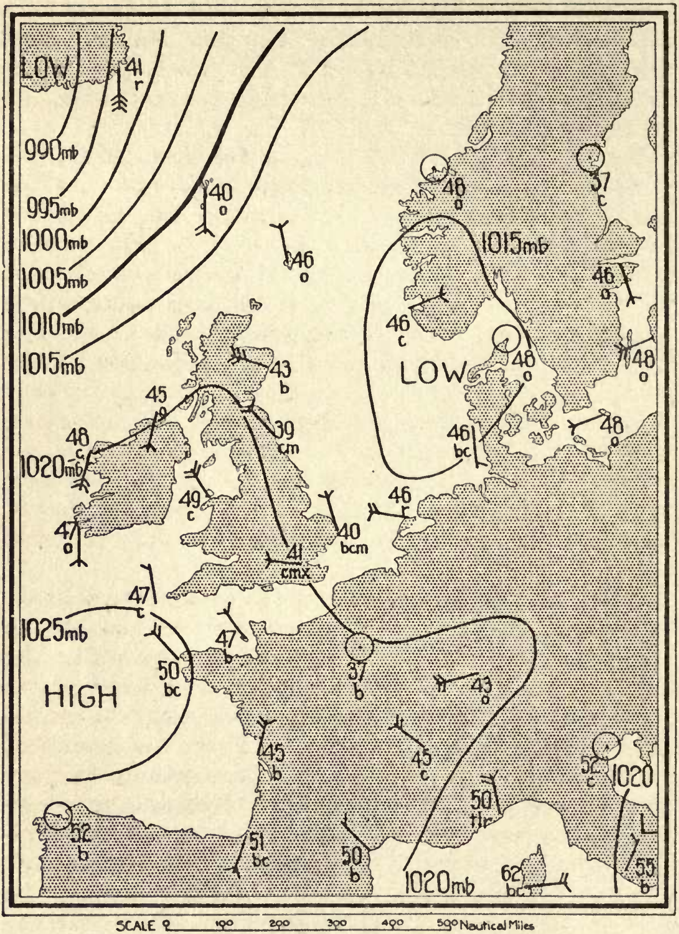
Secondary Depression. 7 a.m., Thursday, 1st January, 1920.

FIG. 9

of the isobars are often associated with very noteworthy weather phenomena in the form of heavy rain or showers. Slight undulations in the isobars—often referred to in reports issued to the newspapers as ill-defined secondaries—are generally an indication of showery conditions.

Fig. 10, the map for 7 a.m. on October 19th, 1917, shows an example of an anticyclonic wedge. As the name implies the isobars form a wedge like projection between two depressions. The wind on the eastern side of the axis of the wedge is north-westerly and may be regarded as belonging to the depression which is passing away, while on the western side of the axis it is from southerly directions, an indication of the oncoming depression. A wedge is usually a very transient distribution, being merely an interlude between two depressions but during its passage the weather is often brilliantly fine as in the cloudless type of anticyclone, though in the example shown, which will be referred to again later (p. 115), this was not so.

With this brief description of the commoner weather types used in meteorological reports we pass on to consider in greater detail the observations which the meteorologist has to piece together in order to arrive at an understanding of the processes going on in the atmosphere which surrounds us. From the nature of the case the meteorologist's inquiries cannot be pursued by the path of experiment. Nature's processes are beyond our control, our line of attack must be by patient organised observation and subsequent grouping of the observations to furnish answers to definite questions. Until recent times such observations have been confined almost exclusively to observations made at the earth's surface, that is to say at the lower boundary of the atmosphere whose processes we are investigating but during the last decade of the



Anticyclonic Wedge, 7am, Friday, 19th, October, 1917.

FIG. 10

past century meteorologists began to attack the problem of organising regular soundings in the free air by means of kites and balloons. Much new and valuable information has thus been acquired and has rendered it possible to apply the test of observation to problems which formerly could only be attacked from the theoretical side.

CHAPTER II

PRESSURE

THE invention of the mercury barometer by Torricelli (1608–1647) may be regarded as the birth of Meteorology as a science. The well known lecture experiment of preparing a model of Torricelli's barometer will be familiar to most readers. A long glass tube is completely filled with mercury, the open end of the tube is then closed with the finger and plunged into a dish containing more of the liquid. On removing the finger the mercury falls, if the tube be long enough, part of the way down the tube leaving a vacuum at the top. The mercury is kept in position in the tube by the pressure which the atmosphere exerts on the mercury in the dish and the height of the top of the mercury column above the surface of the mercury in the dish will be of the order of 30 inches or 76 centimetres. The length of the column is not invariable, if the pressure exerted by the atmosphere increases the mercury rises in the tube and *vice versa*. The length of the mercury column thus gives a measure of the atmospheric pressure. The mere specification of the length of the column in the crude instrument which is here described would give an incomplete account of the atmospheric pressure. The length of the column depends on other factors as well as on the atmospheric pressure, such as the temperature of the mercury and the value of the acceleration due to gravity which in turn depends on the latitude. Variations due to such causes are relatively small and can be allowed for by calculation if the experiment is made, as it always is in practice, under conditions which differ from the standard conditions that the mercury should be at the temperature of the freezing point of water and that the acceleration due to gravity should have the value which it has in latitude 45°. It is, however, most important that the corrections should be applied, for of all the ordinary meteorological observations that of pressure is almost the only one in which great accuracy is both attainable and necessary. Every reading of a mercury barometer, if it is to be regarded as a contribution to meteorological science, must therefore be corrected. The application of the correction is a simple matter. In practice it resolves itself into looking up in a table the magnitude of the correction corresponding with the temperature read off on a thermometer "attached" to the barometer and adding or subtracting this amount from the reading of the barometer. The correction due to the variation of the acceleration due to gravity from its standard value in latitude 45° is nearly constant for any one place and the corrections for temperature and gravity can be combined into a single table. Appropriate tables are given in most books of instructions for taking meteorological observations. Any correction for actual errors in the graduation of the instrument can be incorporated in the same table.

When suitably corrected the specification of the length of the vertical mercury column thus gives a complete specification of the atmospheric pressure—often referred to conventionally as "the height of the barometer," and so general has this specification become that it is adopted also for graduating aneroid barometers, though these instruments are constructed on an entirely different principle. The essential part of their mechanism is a metal box or chamber which is kept distended by a steel spring and has been hermetically sealed after exhaustion of the air within it. Increase in the pressure of the atmosphere compresses the "vacuum box" and spring inside it and a simple mechanism communicates the motion of the walls of the box to the needle which moves across the dial of the instrument. The graduations on the latter have until recently been in terms of the equivalent lengths of the mercury column in a mercury barometer.

An aneroid barometer is not independent of change of temperature. The shape of the box and spring are liable to variation with a rise or fall of the temperature but the changes are so complicated that they cannot be conveniently allowed for by calculation or in tables of corrections. Instead we rely on the instrument maker to arrange that changes of temperature may affect different parts of the mechanism in opposite manner, thus securing a compensation.. The exactness of this compensation for temperature is one measure of the goodness or badness of the instrument. In practice no correction for temperature is applied to the reading of an aneroid barometer if the instrument be used under the ordinary conditions prevailing at the surface. The compensation for temperature is rarely sufficiently complete if the instrument is to be used under conditions which expose it to very extreme temperatures, as in balloon ascents to great heights. For accurate work under such conditions, each instrument must be individually tested by subjecting it experimentally to known changes of temperature and observing the change of reading obtained thereby. Suitable corrections for the instrument can then be worked out. Neither is correction for latitude needed, as the amount of compression of the metal spring is not affected by a variation in the intensity of gravitational attraction.

The indication of an aneroid barometer may alter in consequence of changes in the elastic qualities of the spring and vacuum boxes, and such changes are specially liable to occur if the instrument has been recently subjected to rapid and violent changes of pressure. In inferior instruments such changes of zero, due to internal changes are often very considerable and render the readings quite unsuitable for scientific purposes. No aneroid should be used for such purposes which has not been carefully tested as to the adequacy of its compensation for temperature and the smallness of the "creep" due to internal changes. The National Physical Laboratory tests and certifies instruments. Even with a certified instrument no opportunity should be lost of comparison of readings with a mercury barometer which is known to be accurate. For these reasons the aneroid barometer is not used at land stations, but it is much used at sea on account of its great convenience and gives useful results if the precautions suggested above are borne in mind. Measurements of pressure in the free air, made from balloons or kites must of necessity be made with aneroid barometers. The mercury barometer is quite unsuitable for such purposes, except perhaps for readings taken in manned balloons.

The specification of atmospheric pressure in terms of the length of a column of mercury has had one unfortunate result, namely that each country has adopted its own unit of length for the purpose. The older units such as the French "lines" have now become obsolete and we have to deal in practice only with length measurements expressed in millimetres or inches, but the inconvenience of conversion before comparison between readings from neighbouring countries is possible, has been serious. Moreover pressure is not a length and its specification in terms of length is inappropriate. As observations at great altitudes in the free atmosphere multiplied, the inappropriateness became more and more apparent and the specification in terms of length was presently abandoned for upper air work. In its place the specification of pressure in terms of the *millibar*, a unit derived from the unit of pressure on the centimetre-gramme-second system of units has been generally adopted in all meteorological publications which deal with the upper air. It is impossible to maintain a distinction between scientific observations made in connexion with the investigation in the upper air and those made for general meteorological purposes, and since May 1914 the millibar has been used for all observations of pressure in official meteorological work in this country. Its adoption holds out a prospect of an ultimate agreement among the meteorologists of all countries. Millibars have already been adopted partially or entirely by the meteorological services of Spain, France and Italy and of the United States of America.

The system of units based on the primary units of the centimetre for length, the gramme for mass and the second for time has been adopted by all countries in the sciences of electricity and magnetism. The unit of pressure on that system is the dyne per square centimetre. In accordance with well-known principles of hydrostatics the pressure due to a column of liquid of height h centimetres and density D grammes per cubic centimetre is hDg dynes per square centimetre, where g is the acceleration due to gravity.

Thus the pressure due to the column of mercury 76 centimetres long in a mercury barometer, in other words the pressure of the atmosphere, is $76 \times 13{\cdot}6 \times 981 = 1{,}013{,}200$ dynes per square centimetre or approximately one million dynes per square centimetre. The name Bar has been coined to denote a pressure of one million dynes per square centimetre.

The *millibar*, the unit adopted for practical purposes by meteorologists is the thousandth part of the Bar. Thus the normal pressure of the atmosphere, represented by a mercury column 760 millimetres long under standard conditions of temperature and gravity is 1,013·2 millibars. This value has been adopted by physicists as the normal in the specification of physical quantities which vary in magnitude with varying pressure. It is approximately the average value at sea level for North-western Europe. Over the British Isles the average value varies between 1,009·3 millibars in Shetland and 1015·7 millibars in Jersey.

The ordinary variations of pressure at sea level met with in the British Isles lie between 940 and 1,050 millibars though both these limits have been exceeded. It is of some interest to see how the departures from normal group themselves. The following frequency table based on observations at Kew, Aberdeen and Valencia for the ten years 1901-1910 gives some particulars. For pressure conditions below the average we may expect readings to be

between 940 and 950 millibars on 1 day in 15 years
,, 950 ,, 960 ,, ,, 1 ,, ,, 4 ,,
,, 960 ,, 970 ,, ,, 3 days in 2 years
,, 970 ,, 980 ,, ,, 6 ,, per year
,, 980 ,, 990 ,, ,, 20 ,, ,, ,,
,, 990 ,, 1000 ,, ,, 46 ,, ,, ,,

For pressure conditions above the normal we may expect readings to be

between 1020 and 1030 millibars on 122 days per year
,, 1030 ,, 1040 ,, ,, 33 ,, ,, ,,
,, 1040 ,, 1050 ,, ,, 2 ,, ,, ,,
above 1050 millibars on 1 day in 4 years.

The following exceptional values are extracted from the Meteorological Calendar issued by the Meteorological Office:

January 26th, 1884, 926·5 millibars, 27·33 inches at Ochtertyre.
January 10th, 1913, 929 millibars, 27·44 inches, s.s. *Celtic* in Lat. 50° N. Long. 29° W.
February 5th, 1870, 926 millibars, 27·33 inches, s.s. *Tarifa* in Lat. 51° N. Long. 24° W.
September 22nd, 1885, 917 millibars, 27·135 inches at False Point, Orissa.

September 25th, 1905, 918 millibars, 27·171 inches, s.s. *Pathfinder* in Lat. 12° N. Long. 125° E.
January 9th, 1896, 1054·5 millibars, 31·108 inches at Ochtertyre.
January 8th, 1820, 1052·5 millibars, 31·08 inches, at Gordon Castle.
January 12th, 1915, 1071 millibars, 31·62 inches, at Irkutsk (Siberia).
March 13th, 1900, 1054 millibars, 31·09 inches, s.s. *Lumen* in Lat. 55° N. Long. 24° W., highest reading at sea in North Atlantic.

Reduction to Mean Sea Level. The pressure which the atmosphere exerts is due to the weight of the column of air above the place of observation. It follows that the pressure must vary with altitude, and if we wish to investigate the distribution of pressure over a wide area, say the British Isles, we ought to make all our pressure observations at the same height above Mean Sea Level or if that be impossible corrections must be applied to reduce the readings to a common level. Mean Sea Level is selected as the level for which isobars are drawn on synoptic charts. Unless the barometer happens to be set up at Mean Sea Level, which is actually the case at many of the Dutch stations but is rarely practicable elsewhere, we must add to the observed barometer reading before it is entered on the chart, an amount equal to the pressure which would be exerted by a layer of air of height equal to the height of the station above Sea Level, under appropriate conditions of temperature and pressure.

If p_1 be the pressure which we observe at the top of such a layer and h be its height measured in metres, p_0 the pressure at the bottom, *i.e.* at Sea Level, then the relation between these quantities is

$$\log p_0 - \log p_1 = {\cdot}0148 \frac{h}{\mathrm{T}},$$

where T is the temperature of the layer measured in centigrade degrees on the *absolute scale* of temperature on which the freezing point of water is 273 (see p. 49).

We have thus all the material necessary for calculating p_0 in any given case, and indeed for constructing a table appropriate for the station, showing the amounts to be added to *p*, the observed pressure under varying conditions. It will be seen that the values in such a table depend on T the temperature which we choose to assign to the air. We can obtain the temperature of the air at the place where the observation is made by direct observation and this is the value of T generally selected when picking out from the table the value of the increment appropriate to the particular occasion, but in following this convention we may be introducing disturbing factors. For example the winter temperatures over large continents tend to be very low and the stations in central Europe or Asia are some of them at great altitudes. In thus assigning a very low temperature and consequent high density to the layer of air which we are imagining below the station we may make the increment and consequently the calculated pressure at sea level unduly large. The very high sea level pressures shown over the high region of Siberia or Central Europe in winter are thus to some extent dependent on the convention adopted in reducing the values. In summer when temperature is high, the reverse may be the case and the sea level pressures appear unduly low. The Swiss meteorologists endeavour to get over this difficulty by adopting not the observed temperature, but the average temperature for the month for selecting the appropriate increment for reducing their pressure observations to sea level. In a mountainous country like Switzerland the local temperature readings are often abnormally cold in winter.

Such difficulties would be avoided if we constructed our maps of isobars not for sea level but for some higher level, say the level of the highest station. We should then have to deal in our calculations with an actual layer of air in which the temperature distribution can be investigated by observation. With the multiplication of observations in the free air the day may come when this is practicable; at present the available observational material is too limited, except in cases for which observations are specially made.

The observations of pressure made at the various stations when reduced to a common level give us the means of drawing the isobars. If the isobars are to be correct, a high degree of accuracy in the barometer readings is required. With many meteorological observations the instrument can easily be read to a much higher degree of accuracy than is needed for meteorological study but with the observation of pressure this is not the case. The individual readings must be accurate to a tenth of a millibar. To attain this not only must great care be used by the observers but the instrumental errors must be known and allowed for. The instrumental errors are determined by direct comparison with a standard instrument and in all meteorological services only readings from instruments which have been thus tested and certified are accepted. Moreover the reduction to Mean Sea Level involves an accurate knowledge of the height of the barometer above sea level. A difference of level of 10 metres corresponds with a difference of pressure of 1·2 millibar, so that an accuracy of a tenth of a millibar requires that the height of the barometer cistern above Mean Sea Level should be known to the nearest metre. (The height must be determined by careful levelling from the nearest bench mark shown on the maps of the Ordnance Survey.)

While discussing the question of reduction of barometer readings to Mean Sea Level we may glance at the allied question of the determination of heights by means of barometer observations. The height reached by an aeroplane is generally determined from the reading of an aneroid barometer, indeed these instruments are often graduated to show height and are sometimes called altimeters for that reason. The basis of the determination is the formula which we quoted above, which may also be written in the form

$$\log \frac{p_0}{p_1} = \cdot 0148 \frac{h}{\mathrm{T}}.$$

The aneroid gives values for p_0 and p_1 the pressure at the surface and at the top of the ascent, and for the evaluation of h, the height, we must know T the temperature. The height scales on altimeters in this country graduated in feet are based on a table prepared by Airy which assumes a constant value for T of 50° Fahrenheit, 283 on the absolute scale of temperature. At the time when this table was prepared altimeters were used almost exclusively for determining heights of mountains and 50° F. was accordingly a suitable temperature to select, on the ground that that value represents a rough mean of the values commonly met with at the earth's surface. If the temperature of the air differs from 50° a correction should be applied to the indicated height. If the true temperature is above 50°, the true height is somewhat greater than that shown on the dial and *vice versa*. Airy gives a rule for calculating the amount of the correction.

It ought, however, to be pointed out that the appropriate value for T in the above formula is neither the temperature at the earth's surface, nor the temperature at the top of the ascent, but a generalised value which has in some way to summarise the conditions of temperature prevailing in the layer of air through which the ascent is made. The most satisfactory manner of arriving at the appropriate temperature for an individual case is by direct observation of the temperature of the air with a thermometer at intervals during the ascent. We are then in possession of all the data required for the calculation of the height by the formula given above. If observations of temperature are not available we can form an estimate of the most probable correction to the height as read off on the scale of the altimeter by observing the temperature at the surface and making assumptions as to the rate of change of temperature with height based on results of other observations.

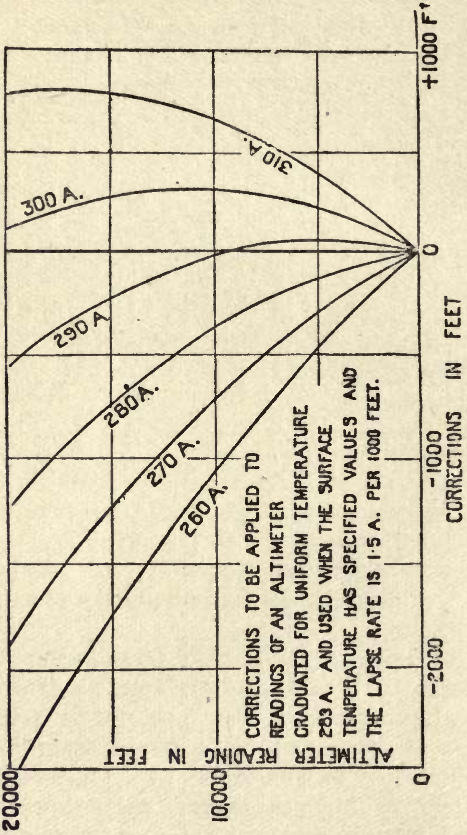


Fig. 11

Unfortunately, as we shall see in a later chapter, the variation of temperature in the first few thousand feet above the ground is often very irregular, so that if we assume that temperature decreases by 1·5 degrees absolute, approximately 3 degrees Fahrenheit, for each 1,000 feet of ascent, which is the average lapse rate or decrease of temperature with height, the assumption may be appreciably in error. It is the assumption most usually made if actual observations of temperature are not available and it leads to the corrections to the altitude scale which can be read off from Fig. 11. It will be seen that the corrections, on this assumption, are small for a surface temperature of 290a, 63° Fahrenheit, but they are appreciable at other temperatures. An altimeter reading of 10,000 feet would be too high by 360 feet under conditions of surface temperature 280a, 45° Fahrenheit, and too low by 310 feet under conditions of surface temperature 300a, 81° Fahrenheit. At more extreme temperatures and greater heights the errors and uncertainties increase rapidly.

CHAPTER III

WIND

THE majority of the observations of wind with which meteorologists have to deal are estimates of its force made on what is known as the Beaufort Scale. Practically all observations made at sea are of this kind and on land also estimates are more common than measurements made with anemometers, and are in some ways to be preferred unless much care is given to the selection of an appropriate site for the instrument. On the Beaufort Scale the values 0 and 12 are assigned respectively to "calm" and "hurricane" and the observer is called upon to assign intermediate values for the winds which he ordinarily encounters. The original specification, drawn up by Admiral Beaufort in 1805, had reference to the man of war of his day. For light winds it specified the speed in knots which such a vessel would make with all canvas set, while strong winds were described by reference to the amount of canvas that could be carried, force 12 being described as "that which no canvas can withstand." The specification has now become obsolete but the tradition lives on among the men of the sea and taking all in all estimates supplied by different observers are surprisingly accordant. For the guidance of observers on land and on modern sailing vessels a revised specification has been drawn up in the Meteorological Office and is reproduced here.

The velocity equivalents given in the last columns of

SPECIFICATION OF THE BEAUFORT SCALE
WITH PROBABLE EQUIVALENTS OF THE NUMBERS OF THE SCALE.

Beaufort Number.	General Description of Wind.	Specification of Beaufort Scale.		Limits of Velocities.*		Beaufort Number.
		Estimating aboard Sailing Vessels.	For use on Land, based on Observations made at Land Stations.	Statute Miles per Hour.	Metres per Second.	
0	Calm	... ...	Calm ; smoke rises vertically	Less than 1	Less than 0·5	0
1	Light air	Sufficient wind for working ship.	Direction of wind shown by smoke drift, but not by wind vanes.	1–3	0·5–1·5	1
2	Slight breeze ...		Wind felt on face ; leaves rustle ; ordinary vane moved by wind.	4–7	2–3	2
3	Gentle breeze ...		Leaves and small twigs in constant motion ; wind extends light flag.	8–12	3·5–5·5	3
4	Moderate breeze	Forces most advantageous for sailing with leading wind and all sail drawing.	Raises dust and loose paper ; small branches are moved.	13–18	6·8	4
5	Fresh breeze ...		Small trees in leaf begin to sway ; wavelets form on inland waters.	19–24	8·5–10·5	5
6	Strong breeze ...	Reduction of sail necessary with leading wind.	Large branches in motion ; whistling heard in telegraph wires ; umbrellas used with difficulty.	25–31	11–14	6
7	High wind ...		Whole trees in motion ; inconvenience felt when walking against wind.	32–38	14·5–17	7
8	Fresh gale ...	Considerable reduction of sail necessary even with wind quartering.	Breaks twigs off trees ; generally impedes progress.	39–46	17·5–20·5	8
9	Strong gale ...		Slight structural damage occurs (chimney pots and slates removed).	47–54	21–24	9
10	Whole gale ...	Close-reefed sail running, or hove to under storm sail.	Seldom experienced inland ; trees uprooted ; considerable structural damage occurs.	55–63	24·5–28	10
11	Storm		Very rarely experienced ; accompanied by widespread damage.	64–75	28·5–33·5	11
12	Hurricane ...	No sail can stand even when running.	... ...	Above 75	34 or above.	12

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* For finding the Beaufort number corresponding to a velocity expressed in miles per hour.

this table are based on a careful comparison of estimates by experienced observers with anemometer measurements from well exposed anemometers.

The comparisons on which these figures are based were, for the most part, with the records of Robinson cup anemometers, in which the wind causes the spin of a suitably poised vane, like a windmill spinning in a horizontal plane. The wind speed is measured by counting the number of rotations in a given time. The record gives the "run" of the wind in a stated interval such as an hour, which is valuable information for many purposes, but it is not suited for disentangling rapid fluctuations of wind.

A more informing record is given by the more modern pressure tube anemometer which has displaced the cup anemometer at many stations. The head of this instrument is relatively light so that it is comparatively easy to obtain a good exposure by mounting it on a tall but light mast. It is so constructed that it measures the difference between the increase of pressure which the wind causes when blowing down the open end of a tube, which is turned by a vane so that is always faces the wind and the decrease of pressure due to suction of the wind as it blows past a row of holes in a closed tube. Fig. 12 shows a facsimile of the record of such an instrument. The anemometer head was in this case exposed on a tall mast 98 feet high at the Observatory at Pyrton Hill and the record reproduced is that from about 9.30 a.m. on December 11th, 1912, to about the same hour on the following day. The trace at the bottom of the diagram is the record of wind direction, the one in the centre that of wind velocity. We see that the wind is by no means steady, but is always fluctuating in velocity. The record was interrupted from about 2.30 to 3.30 p.m. for the more detailed study of these fluctuations. For this purpose the cylinder carrying the record

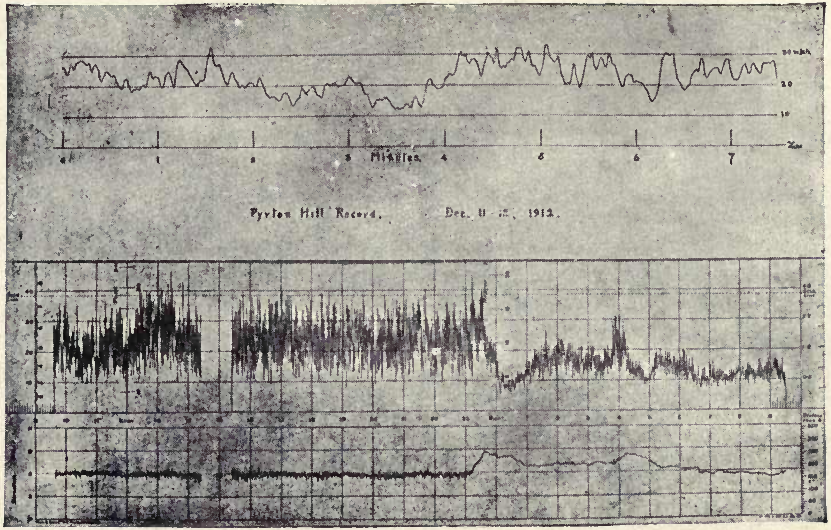


FIG. 12—WIND RECORD BY PRESSURE-TUBE ANEMOMETER,
PYRTON HILL, DEC. 11-12, 1912

sheet was made to revolve more rapidly so as to obtain a record with a very open time scale. The trace at the top of the diagram shows a portion of this enlarged record covering an interval of about 7 minutes. We see that the wind consists of what we may call gusts and lulls following one another in rapid succession, the average time between one gust and the next being about 6 seconds. The record of wind direction reproduced in the lower curve shows that the fluctuations of velocity are associated with small variations in direction.

These variations in the wind, often referred to as its gustiness, are never absent. They are probably at a minimum over the open sea, but even there there is always a certain amount of eddying in the air. When the wind passes over a land surface, hills, cliffs, buildings, trees and other irregularities break up the steady flow and thus the conditions of the exposure require consideration in the interpretation of anemometer records. The traces from an anemometer at a coast station, even if the country behind the station be generally flat and bare, present a very different appearance according as the wind is on or off shore. With the wind direct from the sea the range of velocity between the maxima in the gusts and the minima in the lulls is markedly less than when the wind comes from over the land. Even at inland stations the surroundings are rarely symmetrical and so the gustiness shown by the anemometer trace depends on the wind direction. The record reproduced in Fig. 12 shows a shift of wind from south to south-west between 11 p.m. and midnight, and though there is no very conspicuous change in the mean velocity, the width of the trace is considerably reduced.

As might be expected, the range between the maxima and minima increases with the velocity of the wind. It is roughly proportional to it and as a numerical measure of the gustiness of different exposures we may take the ratio of the range of the gusts, the width of the ribbon of the anemometer trace, to the mean velocity as a "gustiness factor."

Even in a flat country trees and similar objects have a marked influence in breaking up the wind; anemometers exposed in well-wooded country invariably have high gustiness factors even if the heads of the instruments be raised well above the tree tops.

The following figures apply to a number of typical situations.

Southport (Marsh side) flat treeless marsh by the sea ..	0·3
Scilly Isles flat top of a hill	0·5
Shoeburyness flat sand by sea E.N.E. wind from sea ..	0·3
,, ,, ,, W. wind over land ..	0·8
Kew Observatory Roof (Park with large trees)	1·0
Dyce (Aberdeen) Anemometer mast projects 15 feet about tree tops	1·3

On this scale of measurement the gustiness at Dyce, over trees is more than four times that over the marshes at Southport. We may also note the conspicuous difference in the gustiness at Shoeburyness according as the wind blows from over the sea or over the land.

Experiments with anemometer heads exposed at different levels show that the gustiness decreases with height but the influence of the surface conditions in modifying the air flow extends far above heights at which anemometer measurements can be made.

From these considerations it is evident that caution is necessary when specifying a wind measurement by anemometer in terms of velocity. We may ask ourselves Should we quote the average maximum of the gusts or the average minimum in the lulls or the mean of the two ? A good case could be made out for the first on the ground that the most obvious effects of the wind are those produced by the gusts, but in practice the mean is generally specified as the comparisons with Beaufort estimates, which have led to the adopted velocity equivalents for the Beaufort numbers given on p. 37, are with mean velocities derived from cup anemometers. A wind which might be quite correctly set down as "Beaufort 5" from its general effects might give appreciably different velocity values on two anemometers, if the conditions of exposure differed materially. The observer in making his estimate has one great advantage over the anemometer in that he can frame his estimates from the effects which he observes on objects at some distance from his actual position; the anemometer can only record the air flow at the head of the instrument. However the wind is measured there can be no precise connexion between the wind near the surface and the wind in the free atmosphere beyond the range of surface disturbances.

Much information regarding the latter has been accumulated in recent years. In the days before the systematic exploration of the upper air with kites and balloons was taken in hand, information as to the motion of the air in the strata above those which can be made the subject of experiments with anemometers was derived almost exclusively from the observation of the drift of clouds. More recently our knowledge on this subject has been greatly extended by direct observations obtained from records of anemometers, generally of the cup type, hoisted aloft by kites or tethered balloons, and from the records of pilot balloons. The War gave a great impetus to observations with pilot balloons as the readiest means of obtaining the information about upper currents which was required both by aviators and by the artillery. Small indiarubber balloons filled with hydrogen are liberated and allowed to drift with the wind. They are watched through the telescope of a theodolite of special construction until they either burst or are lost to view in cloud or distance, the azimuth and elevation of the theodolite being noted at short intervals. It is preferable to use two theodolites separated by a base line of considerable length, for the available readings then suffice to fix the position of the balloon in space at any observation and hence to lay down its path, from which the direction and velocity of its motion at any level can be determined. If only one theodolite is available the results can only be worked out if the rate of ascent of the balloon is known. In still air the rate of ascent of a balloon of this kind is approximately constant provided that there is no escape of hydrogen due to leakage, and its magnitude can be calculated with reasonable accuracy from the "free lift" of the balloon and its "dead weight" and indeed in practice the free lift is generally adjusted to give a definite vertical velocity, 500 feet per minute being a convenient value. The height of the balloon is then 500 feet at the end of the first minute after liberation, 1,000 feet at the end of the second, 1,500 at the end of the third and so on. If the height be calculated in this way the elevations and azimuths read off on a single theodolite give all the data required to calculate the positions of the balloon at the end of each minute and from these the direction and velocity of its motion can be computed. The method assumes that there are no up or down currents in the atmosphere to modify the rate of ascent. If there are such vertical currents, errors, which may be of considerable magnitude, are introduced into the results. An upward current has the effect of decreasing the apparent horizontal velocity and *vice versa*. An escape of hydrogen through leakage reduces the rate of ascent of the balloon, and increases the apparent horizontal velocity.

Not only are errors due to vertical air currents avoided in the two theodolite method, but we can use the observations to obtain information regarding the magnitude of the vertical component of the motion.

The theodolite readings by themselves give sufficient data for calculating the true position of the balloon at any moment. The difference between the actual rate of rising thus calculated and the normal rate in still air gives information about the vertical component of the air motion. Except in exceptional circumstances this component is small; it appears that it rarely exceeds 2 to 3 metres per second in magnitude. Information obtained with pilot balloons is now being published regularly by the Meteorological Office in a special Upper Air Supplement to the *Daily Weather Report.* The majority of the ascents are made with only one theodolite.

In clear weather soundings with pilot balloons up to 10,000 feet present no difficulty. Under favourable circumstances, clear atmosphere, and light wind, the balloons ordinarily used may be kept in view up to much greater heights, 20,000 or 30,000 feet, and under exceptionally favourable circumstances heights of over 80,000 feet or 25 kilometres have been reached.

Other methods of determining the direction and velocity of the wind at high levels were developed and applied during the War under the stress of military requirements. The smoke from bursting shrapnel is carried along by the wind and serves, as it were, to render the wind visible. Its drift after the explosion can be easily observed and with suitable apparatus the rate of angular displacement of the smoke cloud may be measured. Its approximate height can be calculated by the gunner from the elevation of his gun and the setting of the fuse of the shell. A knowledge of the height and of the rate of angular displacement supplies all the data needed for solving the equations from which the actual motion of the smoke cloud can be calculated. If the drift of the smoke cloud be measured at two stations separated by a sufficiently long base line, the observations of angular displacement by themselves furnish all the data required for solving the equations and the meteorologist may then complete his experiment without the need for ballistic calculations to determine the height of the shell at the moment of bursting. The method affords a very elegant, if somewhat expensive method of determining the magnitude and direction of air motion at high levels. In common with observations by pilot balloons it has the disadvantage that it cannot be used when the sky is obscured by cloud at a lower level than that at which the wind velocity is desired, though it can often be used on days which are too cloudy for satisfactory pilot balloon work, as advantage may be taken of temporary breaks in the cloud canopy.

A method of observation which is not affected by the presence of cloud was developed by the French Military Authorities. It consists in liberating a balloon to which detonating bombs, timed to explode at given intervals are attached. The velocity of sound being known, careful observation of the times of the explosions from several observing posts supply the data needful for calculating the position of the balloon at the time of each explosion and hence the drift of the balloon can be determined as in an ordinary pilot balloon ascent, in which the balloon is observed through a theodolite.

CHAPTER IV

TEMPERATURE

IN addition to the pressure and motion of the air, the meteorologist requires to know its temperature. This is ascertained from the reading of some form of thermometer, most usually a mercury thermometer. In making the measurement it is assumed that the thermometer gives the temperature of the air surrounding it but it must be borne in mind that the reading of the instrument gives only its own temperature and that some care is required to make sure that it is really in thermal equilibrium with the air that surrounds it. For example if the thermometer is exposed to direct sunshine or to the radiation from hot bodies it may itself acquire a temperature which is considerably higher than that of the surrounding air. For ordinary meteorological observations in this country the thermometers are placed in a screen or box with louvred sides made of wood and painted white. The instruments are thus protected from the direct action of the sun while the louvred sides allow of sufficient ventilation to secure mixing of the air in the screen with the outer air. If the exposure is sufficiently open and a breeze is blowing the precautions are adequate but on calm sunny days temperatures obtained in this way are liable to be somewhat in excess of the true air temperature. Different types of screen may give sensibly different readings, and uniformity of practice throughout any network of stations is very important. The specification of the dimensions and other particulars of the standard Stevenson screen drawn up by a Committee of the Royal Meteorological Society will be found in the "Observer's Handbook" issued by the Meteorological Office. The Stevenson screen used is not equally suitable in tropical climates where the sun's radiation is more intense. In India meteorological observers expose their thermometers in huts with open sides, thus securing much more ample ventilation.

In the case of temperatures measured in the upper air care must also be taken to guard against inaccuracies arising from bad exposure. If the thermometer is hoisted in a kite no elaborate precautions are needed as the wind required to raise the kite assures ample ventilation. The same is true of temperatures recorded on aeroplanes which are now available for securing records of air temperature up to considerable heights, but care must of course be taken to make sure that the thermometer is not in a position where it can be affected by hot gases from the engine. Apart from this obvious precaution screening of the bulb from direct sunshine is all that is needed in these two cases, but in the case of temperatures determined in manned free balloons more elaborate precautions are needed. The balloon drifts with, and is at rest with regard to the surrounding air, and hence ventilation is almost absent. The basket and envelope of the balloon may be exposed to very intense solar radiation and are warmed thereby to a temperature far in excess of that of the surrounding air. Many of the older observations of temperature made in manned balloons were quite valueless from failure to guard against the heating of the thermometer by radiation from the balloon. To secure satisfactory results an aspiration thermometer must be used. In this instrument the sensitive part of the thermometer is placed inside a thin metal tube through which a brisk current of air is drawn by a fan driven by clockwork. The instrument must be mounted on a long rod and held well over the side of the basket so that the air sucked through the tubes may not be air which has been in contact with the balloon.

The most interesting records of temperature in the upper air are those for great heights derived from so-called registering balloons, unmanned balloons to which self-registering instruments are attached. At great heights the balloon either bursts or sinks gradually from loss of gas. In either case it ultimately returns to earth and in a high percentage of cases the instruments have been picked up and returned by the finder in accordance with instructions given on the label which is attached to them. Such soundings of the atmosphere have been carried up to great heights by this method. Ascents to 20 kilometres are not uncommon. So far as the author is aware an ascent at Padua for which a height of 37 kilometres is claimed, still holds the record. In such ascents the recording instrument is attached to the balloon by a thread about 100 feet long to ensure that the temperature record shall not be falsified by dragging the instrument through the air which may have been warmed by contact with the balloon. A light metal case shields the instrument from the sun's rays and the ascensional movement of the balloon as a rule suffices to secure adequate ventilation. It is intended that the ascent should be terminated suddenly at great height by the bursting of the balloon but if the rubber is defective there may be gradual loss of hydrogen by leakage and in such cases the balloon may float for some time at great heights. The temperature record is then liable to be falsified by radiation effects but to an experienced observer the appearance of the record shows when this mishap has occurred and there is nothing for it but to reject that part of the observation. To avoid such contretemps ascents are often made at night, or just before sunset, but it is undesirable to limit ascents to the night hours. The conditions prevailing during the day equally require investigation.

Many instruments have been designed for investigations of this kind. That used in this country was devised by W. H. Dines. It mounts a small aneroid barometer and a bi-metallic thermometer. The records of these instruments are scratched with pin points on a strip of copper about the size of a penny piece, and the record is subsequently read by means of a microscope. The whole apparatus weighs only about 2 ounces, including the case which protects it from the sun. A detailed description of the instrument is given in "The Computer's Handbook."

Here we are concerned only with the results. The ascents supply a continuous record of the pressure and temperature from the surface to the highest point of the ascent such that we can assign to any pressure the corresponding temperature. From these the corresponding heights can be calculated, but before considering the details of the process we must consider the units in which the quantities are expressed.

The pressures will be specified in millibars as explained in Chapter II. In this country temperature is ordinarily expressed on the Fahrenheit scale on which 32 represents the freezing point of water and 212 its normal boiling point, the range between the two fixed points being thus 180 degrees. On the Continent the centigrade scale, on which the freezing point is called zero, and the boiling point 100, is in general use. It must be admitted that the Fahrenheit scale is very convenient for the ordinary observations in this country. The degree is conveniently small so that there is generally speaking no need to observe to a fraction of a degree, except when the humidity of the air is determined from thermometer readings, in which case a high degree of precision is needed. All air temperatures ordinarily met with lie within the range 0 to 99 and thus two figures suffice for expressing temperature, a consideration of some practical importance when printing large numbers of figures. But perhaps the greatest advantage over the centigrade scale is that there is no necessity to have to resort to the negative sign to express temperatures below the freezing point. Unfortunately the zero of the Fahrenheit scale is not quite low enough to cover the whole range of temperature met with in all parts of the world. In polar regions and also in regions like Siberia or Canada temperatures below zero Fahrenheit are by no means uncommon. When we turn to the upper air this special advantage of the Fahrenheit scale is quite lost. The temperatures recorded in registering balloon ascents which exceed a height of about 3 miles are generally below zero. When dealing with the upper air meteorologists have therefore adopted the absolute scale of temperature on which the freezing point of water is 273, and the normal boiling point 373. The degree is, therefore, the same size as on the centigrade scale. The absolute scale is adopted, not merely because it avoids the use of negative values completely, but because almost all calculations into which temperature enters which we may desire to make in connexion with the investigation of the upper air require the specification of temperature on the absolute scale. The publication of results according to any other scale would involve conversion of the values before any computation could be undertaken. For example the formula for the determination of height from observation of pressure (see p. 33) requires the specification of temperature on the absolute scale.

To return to the consideration of the working up of the results of an ascent of a registering balloon, so as to determine the heights associated with the values of temperature and pressure deduced from the record.

Let us suppose that from a given ascent we deduce a series of temperature values T_1, T_2, T_3 etc. corresponding with pressure values P_1, P_2, P_3 etc. such that T_1 is the temperature at the height when the pressure was P_1, and so on. If we take the trouble to work up the record for small steps of pressure we are justified in taking the mean of T_2 and T_3 as the temperature appropriate to the layer of air between the levels at which the pressures were P_2 and P_3 respectively. If h be this difference of level measured in metres, our formula becomes

$$\log \frac{P_2}{P_3} = \cdot 0148 \frac{h}{\frac{1}{2}(T_2+T_3).}$$

We can thus evaluate h for this particular pair of values and by repeating the process for each separate pair and adding the values of h so obtained we can determine the height above the ground at which any particular values of P and T were encountered. The process of thus working up the record step by step is somewhat laborious but by the use of suitable methods the labour can be greatly economised, and if temperature values are available the process should be gone through, if accuracy is required at great heights. For great accuracy a correction should be applied also for the varying humidity of the air, but the necessary data for this are not available as a rule. (Particulars of rapid methods of working and for applying the correction for humidity may be found in the "Computer's Handbook.")

CHAPTER V

CLOUDS

WHEN discussing the determination of the direction and velocity of the wind at levels beyond the reach of measurements by anemometers, mention was made of the fact that the earlier information on this subject was derived from observations of cloud drift. The minute drops of water which compose the cloud move with the air in which they are suspended and in most cases we may take it that the apparent motion of a cloud as observed from the ground is the same as the actual motion of the air, in other words, that the drift of the cloud, if we measure it, will give the wind motion at the level of the cloud. This assumption is not invariably justified. There are cases in which there is persistent cloud formation in consequence of the motion of the air at points in the atmosphere where special conditions prevail. The most familiar example is the formation of a cloud in consequence of a damp wind blowing up the side of a hill. In such cases the cloud which is always being formed at a particular level on the hillside, appears stationary to an observer regarding it from below but it would be an obvious error to conclude from the stationary cloud that calm prevailed on the hillside.

The observation of the drift of a cloud does not enable us to determine the direction and speed of its motion unless we also know its height. Observation of the rate of angular displacement made at a single observing point cannot supply the data needed for solving the equations involved. In recent years many measurements of cloud heights have been made incidentally to experiments with kites and balloons, and in quite recent times by direct observation from an aeroplane. The earlier measurements of the heights of cloud were based on the simultaneous observation of the bearing and angular elevation of a cloud made at two observing points separated by a sufficiently long base line. The determination is not always an easy one as if the base line is of adequate length there may be some uncertainty as to the identity of the particular patch of cloud which is being made the subject of observation. Nevertheless, sufficient measurements were made in this way to enable meteorologists to fix approximate values for the height ranges within which the various cloud types occur.

The study of clouds is a most fascinating one. It makes its appeal to the artist as well as to the scientist. To the meteorologist there is the two-fold interest of unravelling the physical processes which give rise to the ever varying forms which come under observation, and determining from the motion of clouds the motion of the upper air in its relation to the phenomena of atmospheric circulation.

The accepted classification of cloud forms is based on the classification of Luke Howard who distinguished three principal cloud types, viz.:

Cirrus type, of fibrous or feathery appearance.

Cumulus type, having rounded tops.

Stratus type, arranged in horizontal layers or sheets.

Types intermediate between these primary forms are indicated by combining the names of the primary types. Successive writers on the subject suggested new sub-types or variants on the principal types and as time went by the need for a standardisation of the nomenclature made itself more and more felt. This task was ultimately undertaken by the International Meteorological Committee which supervised the preparation and publication of an International Cloud Atlas for general guidance. This Atlas distinguishes the Cloud forms as follows:—

1. *Cirrus* (Ci). Detached clouds of delicate or fibrous appearance, often showing a featherlike structure.

2. *Cirro-Stratus* (Ci-St). A thin whitish sheet of cloud, sometimes covering the sky completely and giving it a milky appearance, at other times presenting more or less distinctly a formation like a tangled web.

3. *Cirro-Cumulus* (Ci-Cu). Mackerel Sky. Small globular masses or white flakes so delicate that they cast no shadows or only very slight shadows, arranged in groups or lines.

4. *Alto-Stratus* (A-St). A thick sheet of a grey or bluish colour sometimes forming a compact mass of dark grey colour.

5. *Alto-Cumulus* (A-Cu). Largish globular masses, white or greyish in colour arranged in groups or lines and often so closely packed that their edges appear confused.

6. *Strato-Cumulus* (St-Cu). Large globular masses or rolls of dark cloud frequently covering the whole sky, especially in winter.

7. *Nimbus* (Nb). A thick layer of dark cloud, with rugged edges from which steady rain or snow is usually falling.

8. *Cumulus* (Cu). Thick cloud of which the upper surface is dome shaped while the base is horizontal.

9. *Cumulo-Nimbus* (Cu-Nb). Shower cloud. Heavy masses of cloud rising in the form of mountains, turrets or anvils.

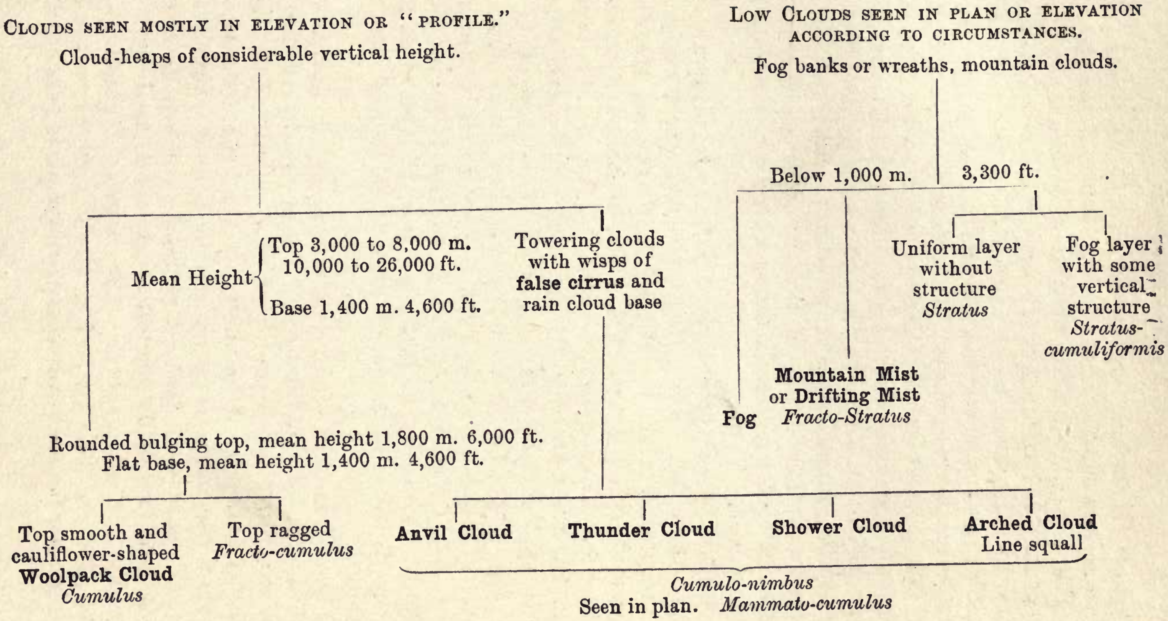
10. *Stratus* (St). A uniform layer of cloud showing no structure.

To indicate that the cloud is broken up, the quali-

GUIDE TO THE IDENTIFICATION OF CLOUD-FORMS.

CLOUDS SEEN MOSTLY IN PLAN

- Stratified in a sheet or continuous layer
  - Mean Height, 9000 m. 30,000 ft.
    - Thin veil without shadows (Halo sheet)
      - Tangled web *Cirro-stratus*
      - Without structure *Cirro-nebula*
  - Thick layer grey or bluish, not rainy (Corona layer) 3,000 m. to 7,000 m. *Alto-stratus.* 10,000 ft. to 23,000 ft.
  - Shapeless Cloud with ragged edges } Below 2,000 m. or 7,000 ft.
    - Rain Cloud *Nimbus*
    - Rain Cloud with vertical extension *Nimbus-cumuliformis*
    - **Scud** *Fracto-nimbus*
- Stratified in separate but ordered cloudlets forming groups (flocks) in layers, lines, or stripes
  - Mean Height 9.000 m. 30,000 ft.
    - Thread or feather clouds **Mares' tails** *Cirrus*
  - Small Flocks (Fr. *Moutons*) in groups or lines, no shadows **Mackerel Sky** *Cirro-cumulus*
  - Largish detached masses with some shadow (Fr. *Gros moutons*) sometimes contiguous in the middle with detached small flakes at the edges *Alto-cumulus*
  - Height { 10,000 ft. to 23,000 ft. 3,000 m. to 7,000 m.
  - Masses with some vertical structure appearing in rolls or waves sometimes covering the whole sky. Below 2,000 m. or 7,000 ft.
    - *Strato-cumulus*
    - **Roll-cumulus**


CLOUDS SEEN MOSTLY IN ELEVATION OR "PROFILE."
Cloud-heaps of considerable vertical height.
Mean Height
Top 3,000 to 8,000 m. 10,000 to 26,000 ft.
Base 1,400 m. 4,600 ft.
Towering clouds with wisps of **false cirrus** and rain cloud base
Rounded bulging top, mean height 1,800 m. 6,000 ft.
Flat base, mean height 1,400 m. 4,600 ft.
Top smooth and cauliflower-shaped **Woolpack Cloud** *Cumulus*
Top ragged *Fracto-cumulus*
Anvil Cloud
Thunder Cloud
Shower Cloud
Arched Cloud Line squall
Cumulo-nimbus
Seen in plan. *Mammato-cumulus*
LOW CLOUDS SEEN IN PLAN OR ELEVATION ACCORDING TO CIRCUMSTANCES.
Fog banks or wreaths, mountain clouds.
Below 1,000 m. 3,300 ft.
Uniform layer without structure *Stratus*
Fog layer with some vertical structure *Stratus-cumuliformis*
Mountain Mist or **Drifting Mist** *Fracto-Stratus*
Fog

fication *fracto* may be applied to these terms, thus *Fracto-Cumulus, Fracto-Nimbus, Fracto-Stratus.*

The diversity and complexity of cloud phenomena is such that a classification of forms which would make it easy to include all cases cannot be drawn up. As a guide to the observer the scheme set out on pp. 54, 55 has been incorporated in "The Observer's Handbook" of the Meteorological Office. It proceeds from the general principle that some cloud forms are seen generally in plan, others are seen generally in elevation or profile, others again may be seen either in plan or elevation. The limits of height between which the different forms generally occur have been worked into the scheme.

A number of typical cloud-forms are illustrated in Figs. 13 to 20. Figs. 13 to 16 are reproductions of photographs taken by Mr. G. A. Clarke at Aberdeen Observatory on February 4, 1909, and show a gradual transition from delicate cirrus to the denser types. Fig. 13, taken at 10 h. 40 m., consists of cirrus mixed with rippled cirro-cumulus. One hour later, at 11 h. 50 m. (Fig. 14), the cirro-cumulus predominates. Fig. 15, taken at 12 h. 5 m., shows the cirro-cumulus becoming coarser and taking on the form of alto-cumulus, while by 12 h. 50 m. (Fig. 16) the type has changed to a mixture of alto-cumulus and strato-cumulus.

Figs. 17 and 18 show clouds arranged in horizontal layers—*i.e.*, of stratus type. These clouds are not of the structureless character of pure stratus, but show a certain amount of breaking up into irregular masses, so that the name strato-cumulus is more applicable. Fig. 17 is a type of strato-cumulus in which an arrangement into long rolls stretching across a great part of the sky is well marked. The name roll-cumulus is often given to this type. Fig. 18 shows strato-cumulus clouds seen from above; the photograph is taken from

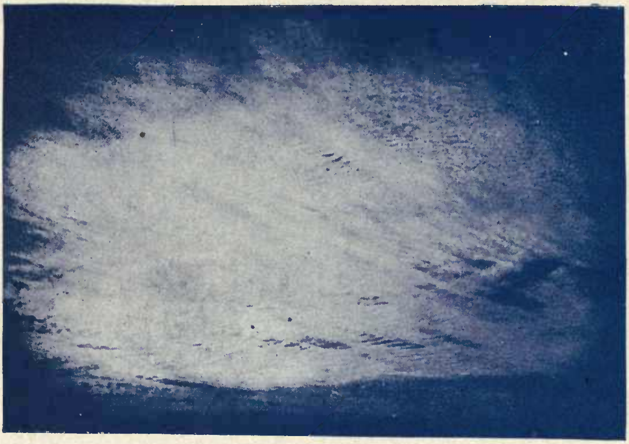


FIG. 13—CIRRUS AND RIPPLED CIRRO-CUMULUS 10 H. 40 M.

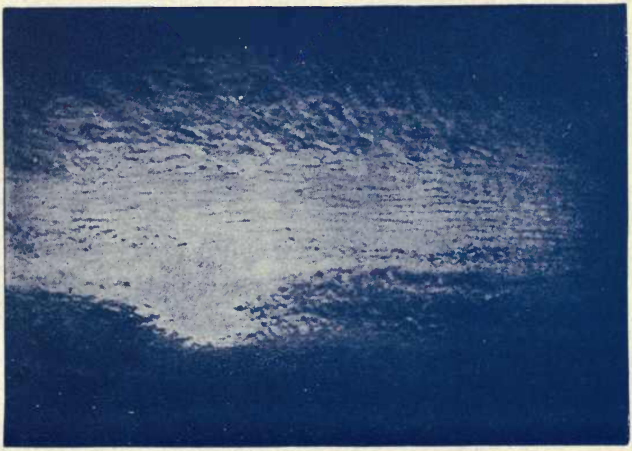


FIG. 14—RIPPLED CIRRO CUMULUS 11 H. 50 M.

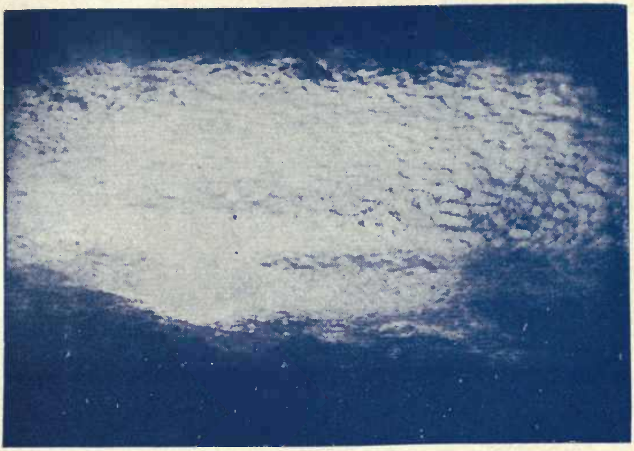


FIG. 15—CIRRO-CUMULUS BECOMING ALTO-CUMULUS 12 H. 5 M.

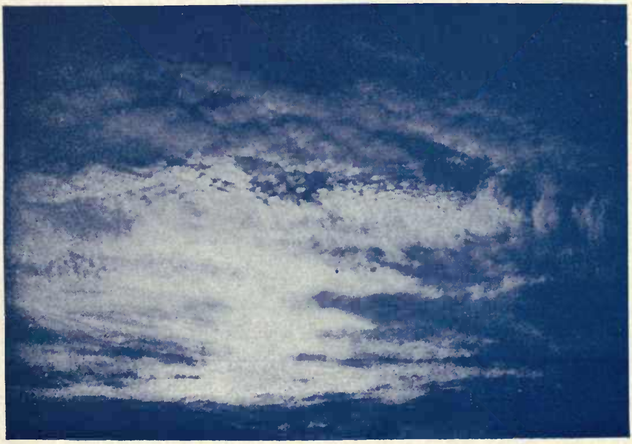


FIG. 16—ALTO-CUMULUS BECOMING STRATO-CUMULUS 12 H. 50 M.

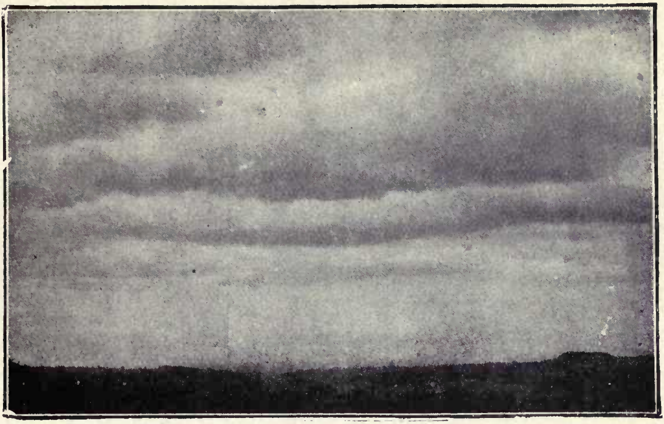


Fig. 17—Strato-Cumulus from below

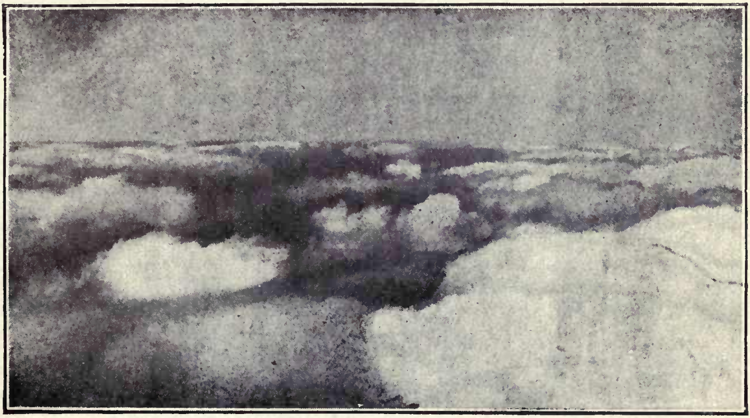


Fig. 18—Strato-Cumulus from an Aeroplane (4,000 Feet)

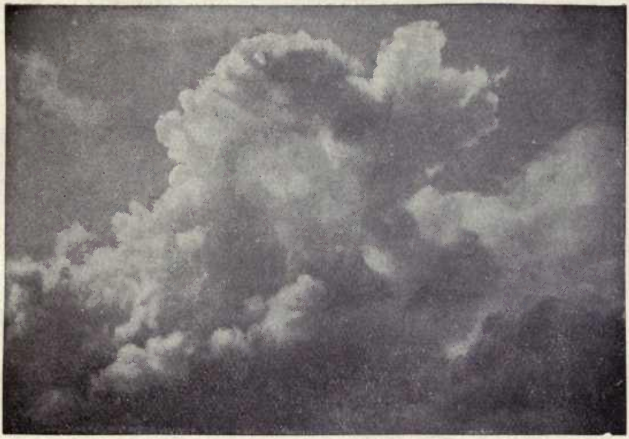


Fig. 19—Cumulo-Nimbus

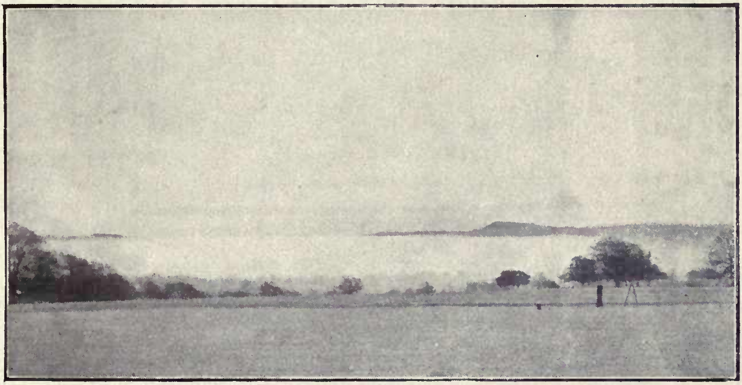


Fig. 20—Valley filled with Fog, 300 Feet Deep, in Early Morning after Still Night

an aeroplane, the height of the upper surface of the cloud being about 4,000 feet. The generally horizontal character of the upper surface is obvious, but the cloud is not uniform and there is a distinct suggestion of arrangement in parallel rolls about this diagram also, though the rolls are somewhat irregular.

Fig. 19 is an illustration of the Cumulus group. The cloud-form suggests obviously that the process of formation of such clouds must be sought in ascending air currents, like steam from a locomotive. The cloud shown in the figure is of cumulo-nimbus type. It is in strong contrast to Fig. 20, which represents what is really a typical structureless stratus cloud. It is actually a photograph of a valley fog seen from above. Note the flat upper surface of the fog through which the tops of the hills on the far side of the valley project.

CHAPTER VI

THE RELATION OF WIND TO THE DISTRIBUTION OF PRESSURE

WE proceed next to consider the interrelation of the various meteorological elements, more particularly that of pressure and wind. In examining weather maps in Chapter I. we noted that the pressure at sea level is rarely uniform over a wide area and we should naturally expect the motion of the air, that is to say the wind, to be closely related to the differences of pressure. At first sight it might even seem natural that the motion should be direct from regions where pressure is high to regions where it is low, but the most casual examination of a weather map shows that this is not the case. We have already noted that as expressed by Buys Ballot's law, the direction of the wind at the surface is more nearly along the isobars than at right angles to them. As regards wind velocity we noted a rough proportionality between the strength of the wind and the distribution of the isobars, the closer the spacing of the isobars, the stronger the wind.

The word gradient is used by surveyors with reference to the slope of the ground to indicate the change of level with distance and is similarly used by meteorologists to indicate the rate of change of a meteorological element over the earth's surface. It is sometimes applied also to the variation of an element with height, but it is better to restrict its use to the case of variation over a horizontal plane. Thus by the pressure gradient we mean the rate at which pressure at the specified level varies over the earth's surface. The gradient is what is known as a vector quantity, *i.e.* it has direction as well as magnitude and the measurement should always be made in the direction of most rapid change. The gradient is inversely proportional to the distance between the isobars, measured at right angles to the isobars. If the latter be drawn for intervals of 5 millibars and the distance between consecutive isobars be x kilometres, we may define the gradient as $\frac{5}{x}$ millibars per kilometre, its direction being perpendicular to the isobars, and directed towards the side of lower pressure.

The theoretical consideration of the connexion between the steady flow of air and the distribution of pressure in the atmosphere of a rotating globe such as our earth may be expressed by the equation

$$\gamma = 2\omega v \mathrm{D} \sin \phi \pm \frac{v^2 \mathrm{D}}{\mathrm{R}} \cot \rho,$$

where γ is the pressure gradient,
v the velocity of the wind,
D the density of the air,
ϕ the latitude of the place of observation,
ρ the angular radius of the path of the wind,
R the radius of the earth,
ω the angular velocity of the earth's rotation.

This equation indicates that there is a balance of forces acting upon the air, or speaking rather loosely that the pressure gradient which tends to push the air towards the region of low pressure is just balanced by the tendency of moving air to turn towards the right (in the Northern Hemisphere), and by the centrifugal force acting outwards from the centre of the curved path. The relation assumes that there is no friction and that the motion is "steady," that is v is not subject to change. The wind thus calculated is called the "*gradient wind.*" We see from this equation that the gradient in its relation to the wind is made up of two components. The second term on the right hand side, sometimes called the *cyclostrophic* component of the gradient, is small if ρ the radius of curvature of the path which the wind describes is large. If ρ is 90° that is if the path is along a great circle, in other words if the path is straight, $cot\ \rho$ is zero, and the cyclostrophic component vanishes. For this special case we may write $\gamma = 2\omega v\mathrm{D} \sin \phi$ and in general in temperate latitudes, unless the path which the air is describing is markedly curved, we shall make no very serious error, in adopting the equation in this simplified form. In most cases the cyclostrophic component $\pm\frac{v^2\mathrm{D}}{\mathrm{R}} \cot \rho$ only figures as a relatively small correction which is moreover not easily calculated as we have not as a rule the data necessary for determining the path and so for evaluating ρ its radius of curvature. The double sign indicates that this cyclostrophic correction may be either positive or negative. It is zero if the path is straight (great circle motion). The positive sign applies in the case of cyclonic motion, *i.e.* motion in a small circle with low pressure at its centre, the negative in the case of anticyclonic motion, with high pressure at its centre.

The term of the gradient wind $2\omega v\mathrm{D} \sin \phi$ is called the *geostrophic* component. It involves the sine of the latitude and is consequently small in low latitudes and zero at the equator. There the cyclostrophic component is the main factor in determining the gradient wind; it may be very large in tornadoes and tropical hurricanes, in which the air is moving rapidly in relatively small circular whirls. The geostrophic component of the gradient gains in relative importance as higher latitudes are reached.

The general rule known as Buys Ballot's law, which we met with in our brief survey of weather maps, was based on the comparison of observed winds with the pressure distribution as summarised by the isobars. The gradient wind is the mathematical expression of the same idea. From what has been said in Chapter III. of the influence of the exposure on the surface wind, we could hardly expect an exact relation between the gradient and the surface wind, though a slight acquaintance with maps suffices to show that the winds are strong when the isobars are closely packed and light when they are openly spaced. Even a comparison of measurements of upper winds with the calculated gradient wind presents some difficulty, as we shall see later, in establishing the idea of proportionality, but the conception of the gradient wind is a very valuable one. It teaches us to look upon a map of isobars not merely as a convenient form of summarising the observations of pressure and so classifying maps but as a representation of the air circulation even though no single observation of wind be plotted on the map. The direction of the circulation is along the isobars, the low pressure being on the left hand side in the northern hemisphere, while the velocity of the air can be calculated from the gradient equation and if the isobars are not markedly curved we make no appreciable error in adopting the simplified geostrophic equation in the form

$$v = \frac{\gamma}{2\omega D \sin \phi}.$$

Equation for Geostrophic Wind.*

The Relation between the Earth's Rotation and the Pressure Distribution for Great-Circle-Motion of Air.

The rotation ω of the earth about the polar axis can be resolved into $\omega \sin \phi$ about the vertical at the place where latitude is ϕ and $\omega \cos \phi$ about a line through the earth's centre parallel to the tangent line.

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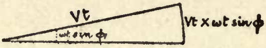
* From the *Meteorological Glossary*, Meteorological Office Publication, No. 225 ii.

The latter produces no effect in deviating an air current any more than the polar rotation does on a current at the equator.

The former corresponds with the rotation of the earth's surface counter-clockwise in the Northern Hemisphere and clockwise in the Southern Hemisphere under the moving air with an angular velocity $\omega \sin \phi$. We therefore regard the surface over which the wind is moving as a flat disc rotating with an angular velocity $\omega \sin \phi$.

By the end of an interval t the air will have travelled Vt, where V is the "wind-velocity," and the earth underneath its new position will be at a distance $Vt \times \omega t \sin \phi$, measured along a small circle,



from its position at the beginning of the time t.

Taking it to be at right angles to the path, in the limit when t is small, the distance the air will appear to have become displaced to the right over the earth is $V\omega t^2 \sin \phi$.

This displacement on the "$\frac{1}{2}$ gt^2" law (since initially there was no transverse velocity) is what would be produced by a transverse acceleration

$$2\omega V \sin \phi.$$

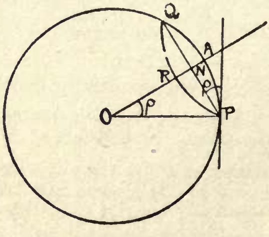
∴ the effect of the earth's rotation is equivalent to an acceleration $2\omega V \sin \phi$, at right angles to the path directed to the right in the Northern Hemisphere, and to the left in the Southern Hemisphere.

In order to keep the air on the great circle, a force corresponding with an equal but oppositely directed acceleration is necessary. This force is supplied by the pressure distribution.

EQUATION FOR CYCLOSTROPHIC WIND.

Force necessary to balance the acceleration of air moving uniformly in a small circle, assuming the earth is not rotating.

Let A be the pole of circle PRQ. Join PQ, cutting the radius OA in N. Acceleration of particle moving uniformly along the small circle with velocity V is $\frac{V^2}{PN}$ along $PN = \frac{V^2}{R \sin \rho}$ where R = radius of earth;



where ρ is the angular radius of the small circle representing the path.

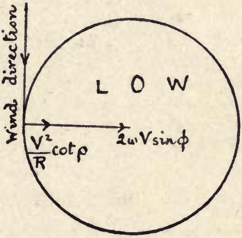
The horizontal component of this acceleration, that is, the component along the tangent at P, is $\frac{V^2 \cos \rho}{R \sin \rho} = \frac{V^2}{R} \cot \rho$.

GENERAL EQUATION CONNECTING PRESSURE GRADIENT, EARTH'S ROTATION, CURVATURE OF PATH OF AIR AND WIND VELOCITY.

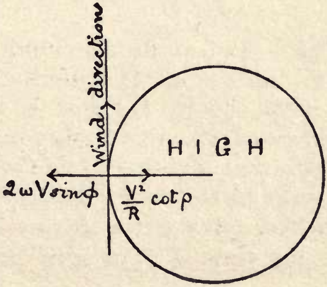
I. *Cyclonic motion.* The force required to keep the air moving on a great circle in spite of the rotation of the earth must be such as to give an acceleration $2\omega V \sin \phi$ directed over the path to the left in the Northern Hemisphere. It must also compensate an acceleration due to the curvature of the path, $V^2 \cot \rho / R$, by a force directed towards the low pressure side of the isobar.

For steady motion these two combined are equivalent to the acceleration due to the gradient of pressure, *i.e.*, $\frac{\gamma}{D}$ where D is the density of the air, and γ the pressure gradient, directed towards the low pressure side.

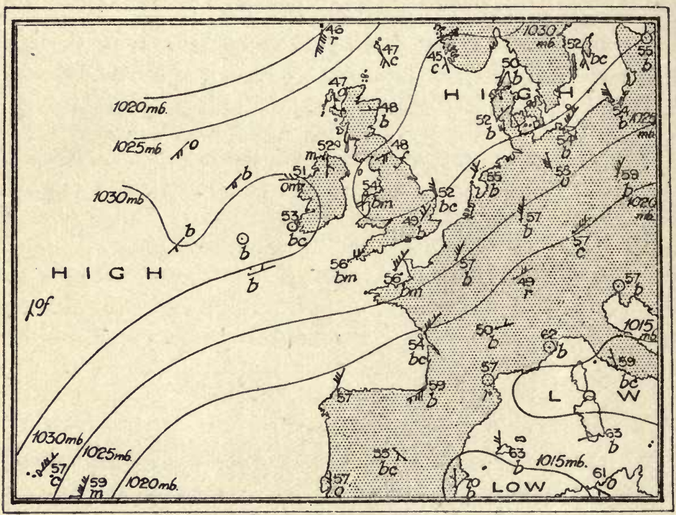
$$\therefore \quad \frac{\gamma}{D} = 2\omega V \sin \phi + \frac{V^2}{R} \cot \rho.$$



II. *Anticyclonic motion.* In this case $2\omega V \sin \phi$ and $\frac{\gamma}{D}$ are directed outwards from the region of high pressure, and the equation becomes $\frac{\gamma}{D} = 2\omega V \sin \phi - \frac{V^2}{R} \cot \rho$.



The conception of regarding the isobars as a measure of the air circulation is especially valuable when we take a survey of a wide area. Fig. 21 shows an example of an Easterly type. Pressure is high over Southern Scandinavia, the British Isles and westward thereof, while over a long strip of the map extending from central Europe to the Azores in Mid-Atlantic, the isobars are parallel curves, representing a great flow of air from East to West. Before we can commit ourselves to the statement that the air is actually transported along the paths indicated by the isobars from Russia to the Azores or beyond, we must satisfy ourselves that the pressure distribution remains constant sufficiently long for the transport to be accomplished. In the case which we have selected the main features of the map remained similar for several days from May 15th till



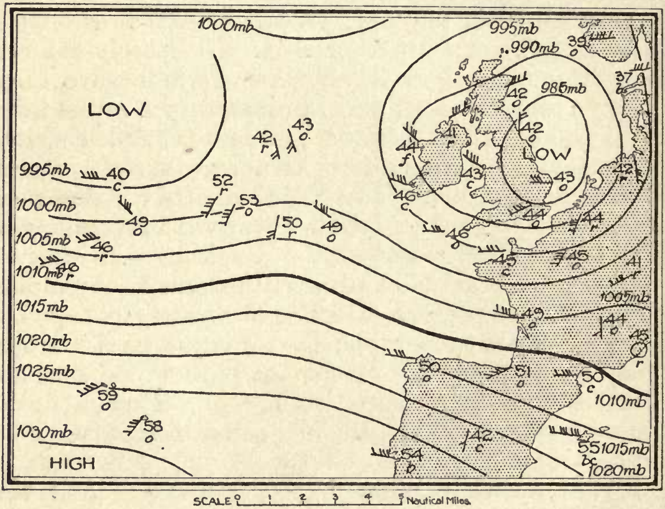
Sunday, 17th MAY, 1914.
7 AM.

FIG. 21

May 18th, and we are, therefore, justified in drawing deductions as to the actual transport of air. The geostrophic velocity over the North of Germany and France at 7 a.m. on May 17th, deduced by measuring the distance between the isobars and applying the equation given above, is about 30 mph. In 24 hours the air would travel 720 miles in the direction indicated by the isobars so that a balloon liberated at Berlin at the epoch of the map might be expected to be located over Brittany 24 hours later and near Longtude 15° W. 24 hours after that. Ultimately the air finds its way to feed the north-east trade wind. In order to maintain such an approximately straight flow of air we see that a pressure gradient is needed which is constantly exerting a force on the air at right angles to its actual motion. If we had control of the elements and wished to arrange for the transport of air along a path such as that described, we should have to steer it by means of a gradient of nicely balanced magnitude which exerts a pressure which is always at right angles to the direction along which the air is to travel. If we made our gradient too strong the path would deviate to the left of the required path, while if we made it too weak, or removed it altogether the air would deviate to the right.

Fig. 22, which shows a map of the North Atlantic Ocean for March 22nd, 1912, gives an example of a gradient steering the air along a course in the opposite direction, from West to East. Pressure is high along the southern border of the map, and there is a wide band of parallel isobars in the middle region of the map indicating a strong flow of air from West to East. Along the northern margin of this wide current of air the flow breaks up into a series of cyclonic depressions which are themselves travelling from West to East in the manner described in our general consideration of the weather map.

The general features of the map in the region south of Latitude 50° N. remained unchanged for several days, so that for that area we may infer a steady flow along the isobars. The geostrophic wind velocity in Latitude 40° N., Longitude 40° W. beyond the western limit of the map as computed from the distance between the isobars is about 40 miles per hour. A run of about 50 hours or about 2 days would therefore carry the air approximately along the course of the isobar for 1,010 millibars, which is shown by a thickened line over the 2,000 odd miles which intervene between Longitude 40° W. and the coast of France.

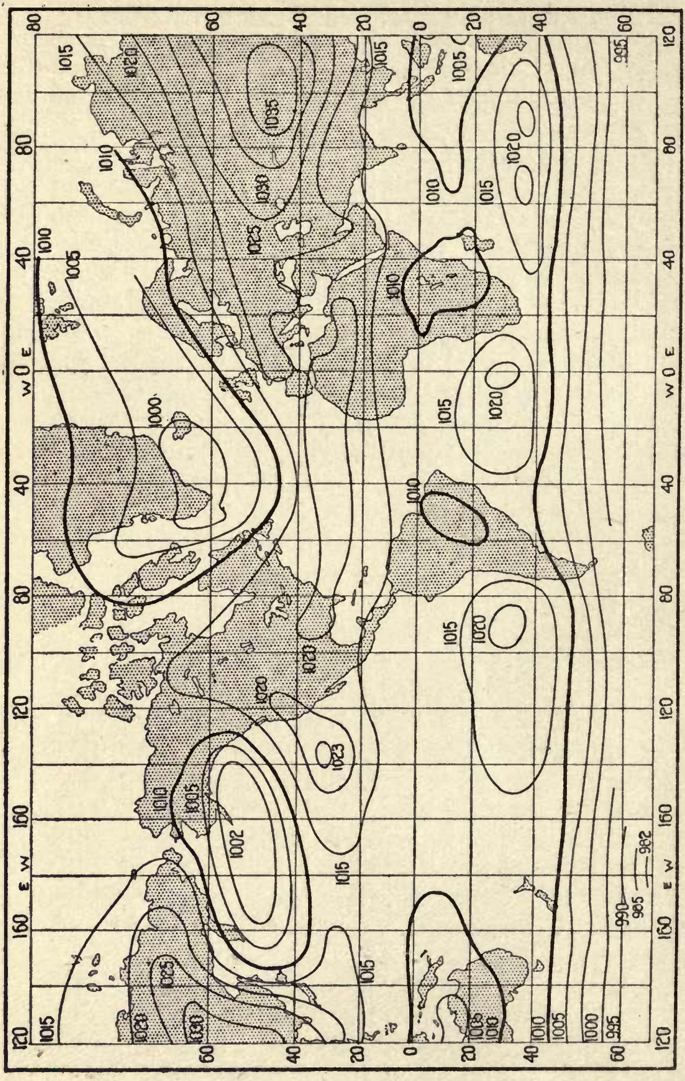


Westerly Type, 7am, Friday, 22nd March, 1912

FIG. 22

Until recently maps of the Atlantic Ocean could only be constructed some time after date when the data supplied by ships' observations had been collected after the arrival of the ships in port. Nowadays wireless telegraphy supplies a more prompt means of collecting the information and the day is probably not far distant when the whole North Atlantic will be meteorologically mapped day by day. A beginning has already been made. A word of warning may not be out of place here lest the reader should carry away a wrong impression as to the normal conditions over the Atlantic. They are by no means as simple as those depicted by the two examples which we have selected. Wide sweeping steady currents covering big areas are the exception rather than the rule. Sometimes there are large anticyclonic areas in which the air motions are slight, also a simple type of conditions, at other times and perhaps most frequently there is a complex system of mixed types, but even then the actual air paths are often much more nearly straight than the momentary isobaric picture shown by an individual map might lead us to expect.

Just as a map of isobars constructed from readings of pressure made at a definite epoch may be regarded as a representation of the flow of air at that epoch, so may a map of isobars constructed from the means of the readings of pressure for a longer period, a month, season or year be regarded as a representation of the resultant air circulation for that period. Fig. 23 shows the average distribution of pressure at sea level over the globe for the month of January. Over the British Isles and surrounding area it shows parallel isobars running from south-west to north-east with low pressure in the neighbourhood of Iceland. The resultant wind over this area is, therefore, represented by a flow from south-west to north-east. This does not mean that in January the wind is always from south-west or even that it is most frequently from that direction. We know that the wind may be from any point of the compass and all we can affirm is that if all the different types of air flow represented by the maps for individual



Average Isobars for the World for January, at Sea Level.

FIG. 23

days are compounded by the parallelogram law, the resultant would be a flow from south-west. In some parts of the world conditions are, however, so steady that the resultant flow is similar to the actual wind met with on most occasions.

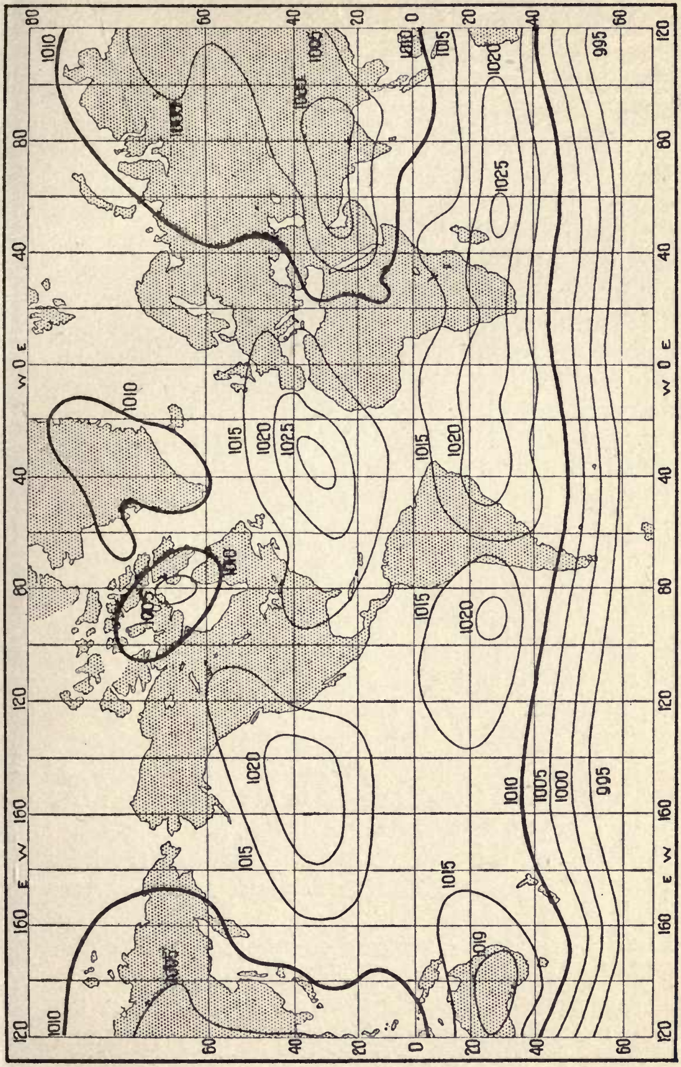
Turning further afield, one of the most conspicuous features of the average distribution of pressure for January in the Northern Hemisphere is a large area of high pressure over Siberia. Pressure is also relatively high over the North American Continent while over the Oceans there are regions of high pressure centred approximately in Latitude 30°. The most conspicuous regions of low pressure are over the Northern parts of the Atlantic and Pacific Oceans, the one centred near Iceland, the other near the Aleutian Islands. It would not be accurate to say that the regions shown as "high" are permanently anticyclonic, all we can say is that anticyclones are frequent there; still less can we describe the regions marked as "low" as permanently occupied by cyclonic depressions, but we are justified in regarding them as areas frequently crossed by depressions. We may connect up the areas shown as "high" by a line which encircles the earth ranging in latitude between 30° and 45°, which marks a sort of ridge of high pressure. This ridge forms a very real line of separation in the winter meteorology of the Northern Hemisphere. To the north of it the resultant winds are from westerly points, south-west over the eastern Atlantic and northern parts of the Eurasian Continent, and also in the Eastern Pacific, north-west in the Labrador region and almost northerly over the east coast of Asia.

Southward of the ridge the resultant flow is from some easterly point. On the Oceans the winds at the surface are from north-east and form the well-known Trade Winds. There we have one of the regions referred to above in which the circulation is remarkably steady. Day by day the same steady north-east wind is met with, with hardly any variation. Over the land areas conditions are more complicated.

In the Southern Hemisphere, consisting as it does mainly of sea, conditions are on the whole simpler. The main features of the distribution found in the north are repeated, but we must remember that Buys Ballot's law is now reversed, low pressure to the right of the course of the wind. Over the Oceans there is a similar ridge of high pressure, approximately in Latitude 35°. North of the ridge the circulation is generally speaking from the east, the south-east Trade wind of the South Atlantic being a very marked feature. The equatorial region is one of low pressure. The equator itself is at this season actually a locus of minimum pressure over the Atlantic and Pacific Oceans. Over the Continents and also the Indian Ocean the locus of lowest pressure is south of the line.

On the southern side of the high pressure ridge we find a belt of parallel isobars, with pressure falling off rapidly polewards. Here in the latitude of the roaring forties strong winds from west predominate. As in the Trade Wind belts, conditions are remarkably steady and the average conditions approximate much more nearly to the actual conditions of any one day than they do in corresponding latitudes of the Northern Hemisphere.

South polar observations suggest that if data were available to extend the map further south we should find a further region of high pressure over the Antarctic Continent. We should therefore, expect to find a southern limit to the region of west winds, beyond which the resultant circulation should again be from the East. In the Northern Hemisphere, there is evidence of such an easterly circulation over the Con-



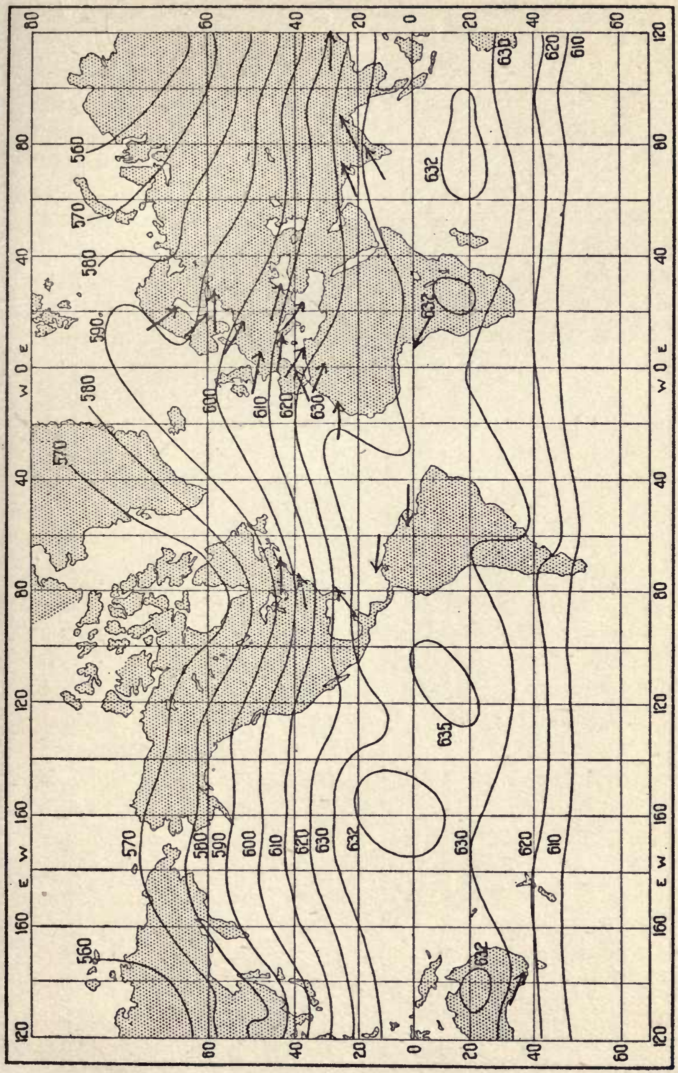
Average Isobars for the World for July at Sea Level.

FIG. 24

tinent of Greenland and also over Alaska and the adjacent part of Siberia.

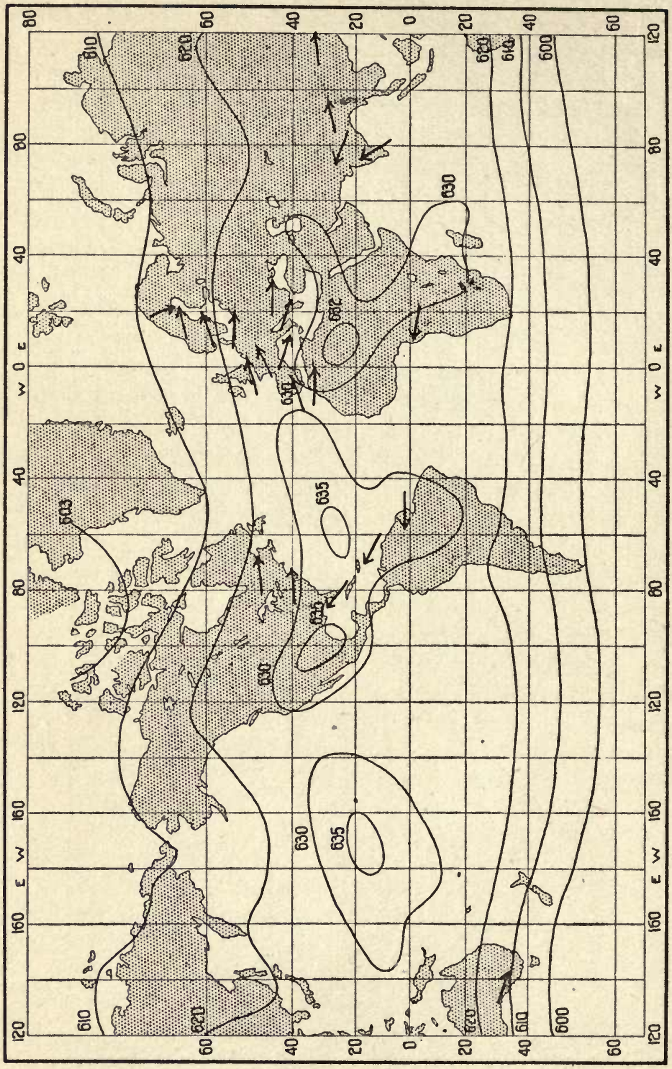
If we turn now to Fig. 24, which represents the distribution in July, the southern winter, we notice that the high pressure ridge of the Southern Hemisphere has become intensified and has shifted somewhat to the north. The westerly circulation on its southern side consequently also extends further to the northward. The equatorial belt of low pressure is generally somewhat north of the equator and in the Indian area is actually displaced as far north as Latitude 30° N. In the Northern Hemisphere the ridge of high pressure extending all round the earth is less conspicuous but the areas of high pressure on the Ocean have increased in intensity and moved to the north. On the northern side we can still identify a circulation with a resultant flow from west but over the Continents conditions are complex and rather indefinite.

The general picture which we get of the distribution of average pressure at sea level both in summer and in winter is that of two belts of high pressure encircling the globe between Latitude 30° and 40° and an equatorial belt of low pressure. The whole system of belts shows a seasonal shift in latitude being furthest north in the northern summer and *vice versa*. In addition there would appear to be polar caps of high pressure. Corresponding with this pressure distribution there is a wind circulation from the east in the equatorial belt between the two high pressure belts, two regions of circulation from the west on the polar side of the high pressure belts and in either hemisphere a polar circulation from the east. The irregular distribution of land and sea introduces many complications into this simplified scheme. When considering the actual maps of average pressure distributions at sea level over the globe, it must be borne in mind that in the continental areas many of the observed pressure values have had



Average Isobars for the World for January at 4,000 metres.

FIG. 25



Average Isobars for the World for July at 4,000 metres.

FIG. 26

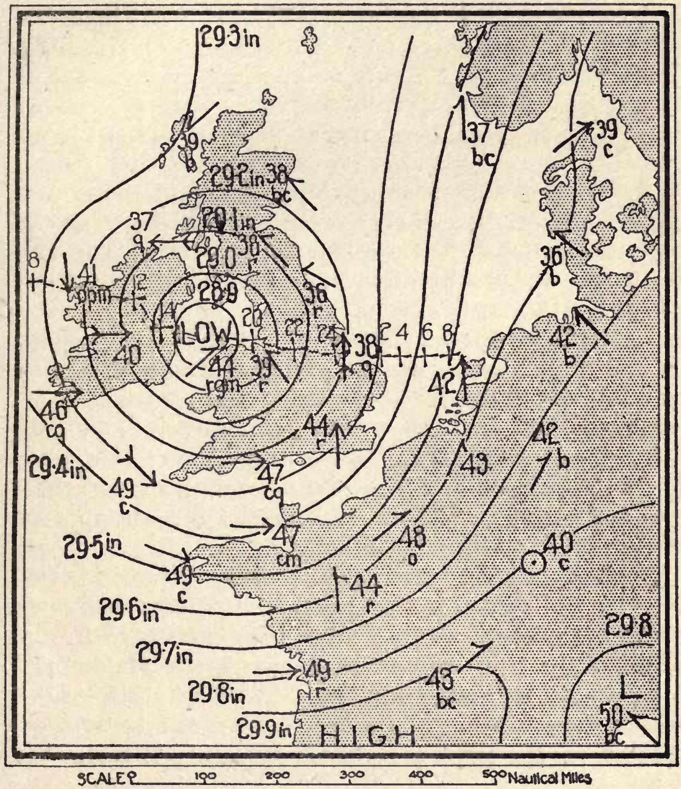
to be reduced to sea level from very considerable altitudes and that in making the reduction the observed temperature values have been used. The isobars for sea level which are shown over Siberia do not correspond with any actual flow of air. At the level for which they are drawn there is solid earth and not moving air. In order to get a satisfactory representation of the resultant flow of air over the Siberian Continent we should draw our pressure map for the average level of the ground, or perhaps better still for the level of the highest station and not for sea level. Teisserenc de Bort has calculated the necessary data for the construction of maps of mean pressure for January and July for the level of 4,000 metres, a level distinctly greater than the general level of the Siberian Continent, but only about half of the height of Mount Everest. The results are shown in Figs. 25 and 26. The general picture is appreciably different from that at sea level. The equatorial belt now appears as a locus of high pressure and pressure falls off on either side towards the poles. The associated resultant wind circulation is an equatorial belt of calms with a circulation from west to east extending from the equatorial belt to the poles. The arrows shown on the maps represent the resultant motion of the upper clouds. The general drift which they reveal is in good agreement with the isobars for the 4,000 metre level.

CHAPTER VII

TRAJECTORIES OF SURFACE AIR CURRENTS

WE have seen that the isobars give a picture of the air circulation at the epoch for which a weather map is drawn but if we wish to discover the paths along which the air actually travels it is unsafe to argue from a map for a single epoch.

The statement is often made that in a circular depression the air at the surface must be approaching the cyclonic centre along a spiral path because it is moving at each point in accordance with Buys Ballot's law, *i.e.* with slight incurvature from the isobars to the side of lower pressure. Such statement leaves out of account the fact that the isobaric picture is itself changing, and before any appreciable part of the supposed spiral can be traced out the isobars may have changed appreciably. For example in the circular cyclone of March 24th, 1902, for which Fig. 27 shows the distribution at 6 p.m. the wind near London, as measured by the anemometer at Kew was about 30 miles per hour from south. Supposing things had remained unchanged for the ensuing 12 hours the distance run through would have been 360 miles and if we lay off that distance along a spiral curve starting in London and maintaining a generally similar incurvature to the isobars represented on the map, we might deduce that the air over London on the evening of March 24th would have arrived as an east wind somewhere near the Mull of Galloway on the following morning. But by that time the centre of the depression was no longer over the Irish Sea, but off the Dutch coast and the wind over South-west Scotland was



Map for 18h(6pm) 24th March, 1902.

FIG. 27

north-westerly and not easterly. Clearly in this case the spiral curve can bear no relation to the actual trajectory of the air.

We can only use a map in this simple manner to determine air trajectories if successive maps are so similar to one another that the lines of flow which they suggest are practically identical. In the case of parallel isobars considered on pp. 64–66 this condition was approximately fulfilled; but in the case of travelling depressions it is not and if we wish to arrive at a true conception of the trajectories of air in such systems we must map the conditions more frequently so that each individual map need only be used to determine the air flow during a time so short that the conditions do not change materially. The normal interval between the successive maps of the public weather service in this country is six or seven hours, which is too long for this purpose, but many stations are equipped with recording barographs and from the records of these instruments we can obtain sufficient intermediate barometric values to construct maps for every two hours or other short interval. The records from anemometers and self-recording thermometers and rain gauges enable us to fill in appropriate wind observations and temperatures and to indicate a part at any rate of the distribution of rain on the intermediate maps.

A set of two-hourly maps has been constructed in this way for the case of the depression which travelled from west to east across the British Isles on March 24–25, 1902. With the help of these maps the trajectories of surface winds shown in Figs. 28 and 29 have been constructed. The process of working out the trajectories from the two-hourly maps is simple, if somewhat laborious.

A simple way is to work with an outline map on tracing paper which can be superposed on any of the completed weather maps. Let us suppose that at a point where we wish to start a trajectory the first map of the series indicates a wind from south-west of 30 miles an hour. In the interval of two hours for which the first map represents the meteorological conditions the wind would complete a run of 60 miles from south-west to north-east. We therefore draw on the tracing paper in the appropriate position a line corresponding with a distance of 60 miles oriented from south-west to north-east. The tracing paper is then superposed on the second map and an estimate made of the direction and velocity of the wind at the point corresponding with the end of the wind step worked out from the first map. The wind step appropriate to the second map is then worked out and marked on the tracing paper. The paper is then superposed on the third map of the series and the process repeated. By proceeding in this way with successive maps the trajectory is constructed step by step.

The broken line running from west to east across the centre of Figs. 28 and 29 indicates the path of the centre and crosses mark its position at different hours of the day. The hours along the trajectories are numbered from 1 to 24 to avoid confusion between a.m. and p.m. We may identify a set of almost straight trajectories exemplified by the curves marked A, B and C showing air motion from south-west along an almost straight line. These curves meet the line of the path of the depression almost at the same epoch as the centre, in other words they end in the region of indefinite wind at the centre of the depression.

Other trajectories D, E, F, L and N showing air motion from south or south-west cut the line of the path some distance in front of the centre ; after crossing the path they loop round and again cut it behind the centre and then continue along a course from north-west, the nearer to the centre the point where they cut the path, the flatter is the loop. O, P, Q, represent a set of trajectories distinct from these loops. They show approximately straight flow from west to east on the south side of the centre, a necessary consequence

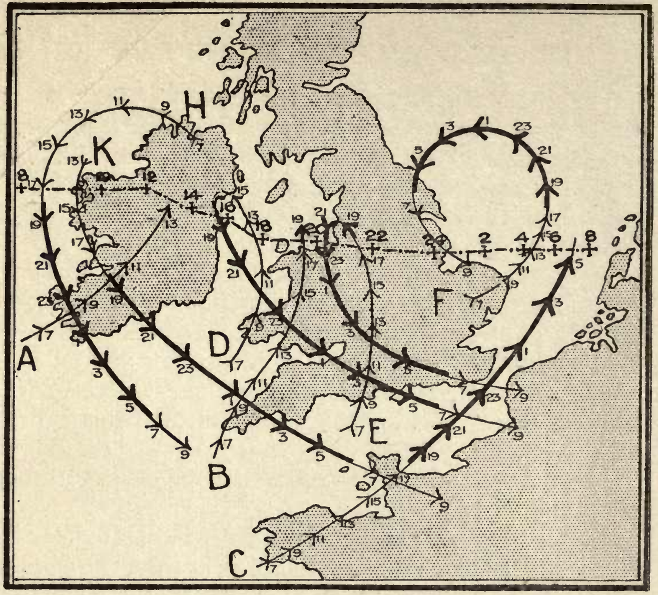

H
K
A
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FIG. 28

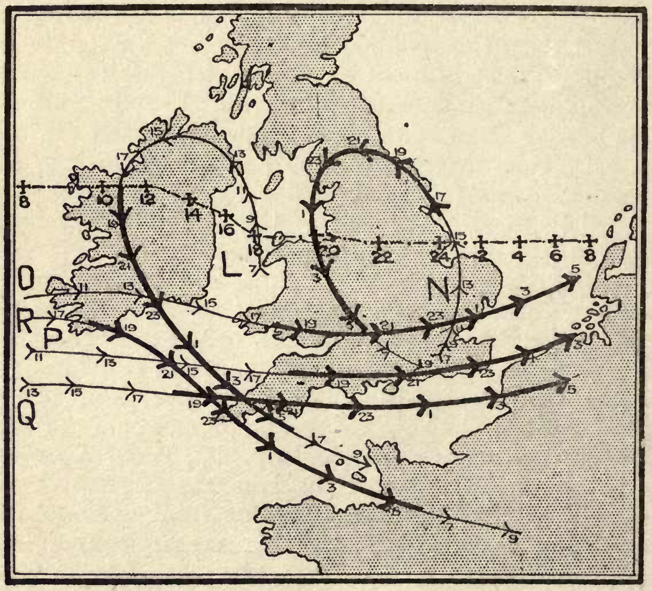

L
N
O
R
P
Q
8
10
12
14
16
18
20
22
24
2
4
6
8

FIG. 29

of the fact that in this depression the rate of motion of the centre was about the same as the velocity of the west wind on the southern side of the centre. These trajectories which represent the flow of the surface winds do not in the least support the idea of a gradual approach of the air towards the centre in all parts of the system along spiral curves. Instead we get the idea of air being fed into the system by the southerly to south-westerly winds in front of the depression, but along the north-westerly trajectories in its rear represented by H, K, and R the air is actually receding from the centre, and is being cast out from the cyclonic system. We may ask ourselves the question what becomes of the air represented by the south-westerly trajectories such as A, B, and C which end in the region of calm or indefinite wind near the centre. We are unable to trace the air further in the surface observations and the conclusion is forced upon us that this air has left the surface and been deflected upwards. Such upward motion is intimately connected with the production of cloud and rain. As air rises its pressure decreases and consequently it increases in volume. The work done in this expansion represents a certain amount of energy, which energy must be taken from the air itself in the form of heat. The air, therefore, gets colder. For air not saturated with moisture the cooling is at the rate of one degree (centigrade) for a rise of 100 metres. If the process be carried far enough the temperature of the air will fall to the dew point and if cooling go beyond this, condensation takes place in the form of cloud and ultimately of rain. This question will be treated more fully in a later chapter. For our present purposes it will suffice if we retain the conception that upward motion of air is associated with cooling and, if persistent, with formation of cloud or rain. Conversely downward motion is associated with warming and if the air contains water drops in suspension, that is in the form of cloud, the warming allows the drops to evaporate and the cloud disappears.

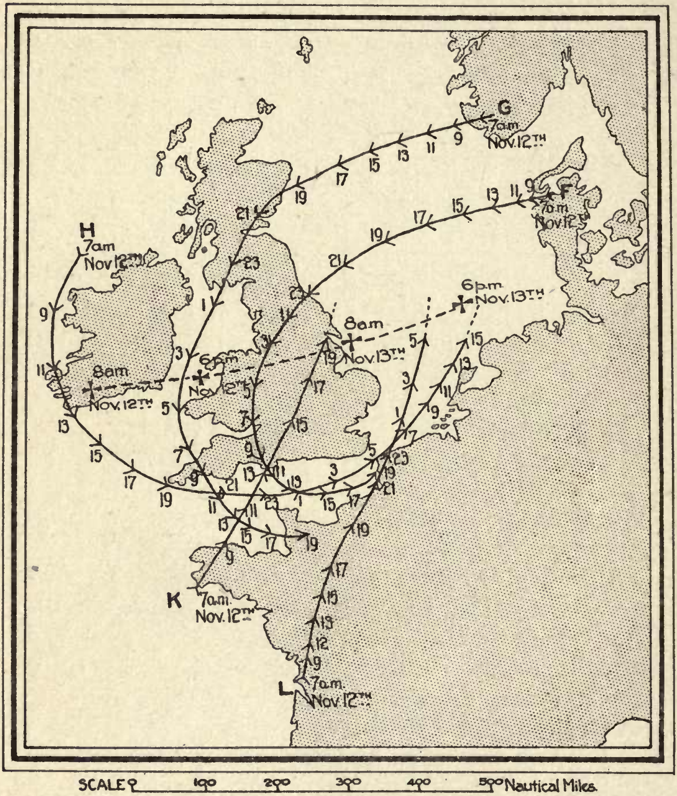
The most obvious examples of the production of rain by the ascent of air are supplied by the forced ascent of air over mountain ranges. Rain thus caused is sometimes spoken of as *orographic*. The heavy rainfall of the Scottish Highlands or of regions like Wales, Dartmoor or the Lake District occurs to us when we survey our own country. Still more striking examples are found on the west coast of India where the south-west monsoon beating up against the high ground of the Indian Peninsula produces an enormous rainfall during the wet season. Over 100 inches of rain fall over the coastal strip on the west coast of the Peninsula from June to October, or in Assam where the hot moist winds from the Bay of Bengal ascend the slopes of the Himalayas. Nature also provides occasional examples of the warming due to the descent of air as in the warm Foehn winds which sweep down the Swiss valleys. The Chinook winds, which descend from the Rocky Mountains over the Western prairies of North America afford another example of such warm down currents.

To return to the consideration of the trajectories of air in a moving depression, it is clear from what has just been said that the area near the centre, where trajectories ending in a region of calm indicate ascent of air, must be one where rain is falling. There must also be appreciable ascent of air, caused by the convergence of the trajectories in front of the centre of the storm, which we have seen is a region of steady rain. If we trace the weather along any of the looped trajectories we find that rain sets in sooner or later in the course of the motion from south, south-east or south-west, and continues until the northernmost point on the loop is reached, or even a little later.

When the motion becomes directed towards the south again the steady rain ceases. Similarly the pressure decreases until the most northerly point is reached and then begins to increase again.

Fig. 30 represents the trajectories of a depression which differed in some noteworthy respects from the case just considered. The depression of November 11–13, 1901, was remarkable for its slow rate of advance, and for the oval shape of the isobars, also for the very heavy rainfall experienced on the northern side of its path. The passage of the depression caused over 3 inches of rainfall over a belt of country, extending over the northern part of Ireland and England, and in some parts of it the amount exceeded 4 inches.

The trajectories of the surface winds are quite different from those of the more rapidly moving circular depression of March 24, 1902. Instead of the single supply of air flowing into the depression from Southerly directions we have here evidence of two very distinct supplies. Trajectories like *K* and *L* show a plentiful intake of air from the south, but in addition we have distinct evidence of a separate supply from the east, on the northern side of the path, represented by trajectories *G* and *F*. This supply from the east is deviated along a course from north in the rear of the depression and ultimately becomes a wind from west on its southern side. We have here a case of two distinct sources from which the depression draws its air and if we look to the temperature conditions of these two supplies we find them conspicuously different. The temperature of the air coming from the south moving along trajectory *K* was between 50° and 55°, whereas in *G* and *F* the temperature was below 40° at the epoch when the air started to cross the North Sea. The trajectories of warm air moving up from the south, *K* and *L*, do not form loops like they did in the case previously considered. They cannot be traced beyond the line of the path, and we are driven to the conclusion that the warm current from the south, when it meets with the colder and therefore denser



Trajectories of November 11th, 1901.

FIG. 30

current from the east rises up above it. In this process the air becomes cooled by expansion and rain results. This is quite consistent with the fact that the heaviest rainfall occurred in the region of east wind on the northern side of the path. The water, though it falls through an east wind was not carried by that wind from Scandinavia across the North Sea. The trajectories show that it came from the south, from the neighbourhood of the Bay of Biscay or beyond. The cold east wind on the northern side of the path of the depression presents an obstacle to the further advance along the surface of the south wind very like a mountain range.

The two cases which we have considered show how fallacious attempts to work out the actual trajectories of air from weather maps, which represent the distribution at a single epoch, may prove, unless we take account of the displacement and other changes taking place in the systems shown. To do so in detail requires maps for very short intervals, but if we select cases that are not too complicated, much can be learned of the transference of air over large areas from maps drawn for longer intervals, and occasionally such trajectories are confirmed by evidence based on quite different considerations. Figs. 33 and 34 represent trajectories for the air over the Atlantic Ocean for the days between February 18 and February 22, 1903, constructed from maps for intervals of 12 hours, two of which, representing the distribution on February 18 and 20 respectively, are reproduced in Figs. 31 and 32.* Throughout the period a constant feature of the distribution of pressure was an anticyclone centred over France or Spain while a series of deep depressions passed over the part of the Atlantic, north of Latitude 50°. The trajectory marked *A* in the diagram is somewhat similar to those marked *O*, *P*, *Q*, of Fig. 29 and represents the flow of air parallel to the path on the southern side of such depressions. The diagram shows us that the air which swept across

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* Figs. 31 to 35 are reproduced from Vol. XXX. of the *Quarterly Journal* of the Royal Meteorological Society, by kind permission of the Council.

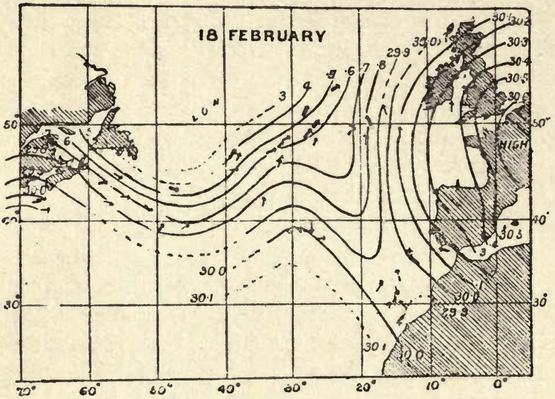

FIG. 31

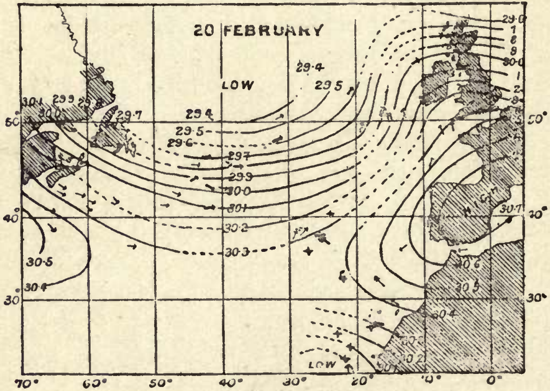

FIG. 32

the British Isles on the morning of February 22nd was derived from two distinct sources. The supply to the northern part of the country came from the western part of the Atlantic, its course can be traced back beyond the 40th meridian (trajectories *A*, *B*, *E*,

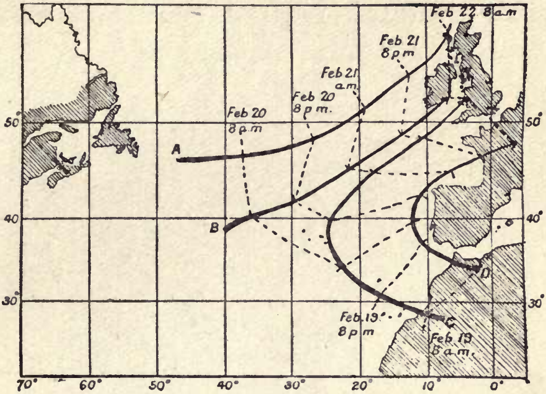

FIG. 33

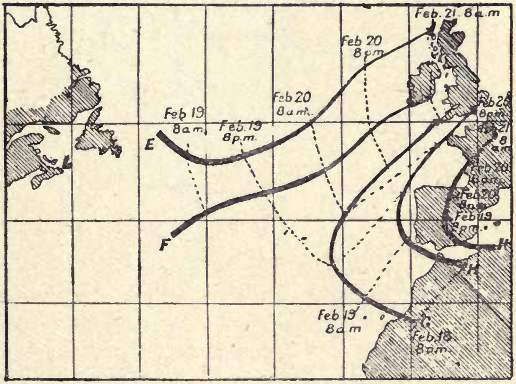

FIG. 34

F). On the other hand the air reaching the south of England made its way round the edge of the anticyclone over Spain and can be traced back to

North Africa (trajectories *C*, *D*, *G*, *H*). The general accuracy of the trajectories was confirmed in a striking manner by a remarkable "blood rain" experienced widely in Southern England and elsewhere. The rain which fell on February 21st and 22nd was discoloured by a curious red deposit, which on examination was identified as probably being of African origin. Its mineralogical characteristics were very similar to those of deposits collected during dust storms off the

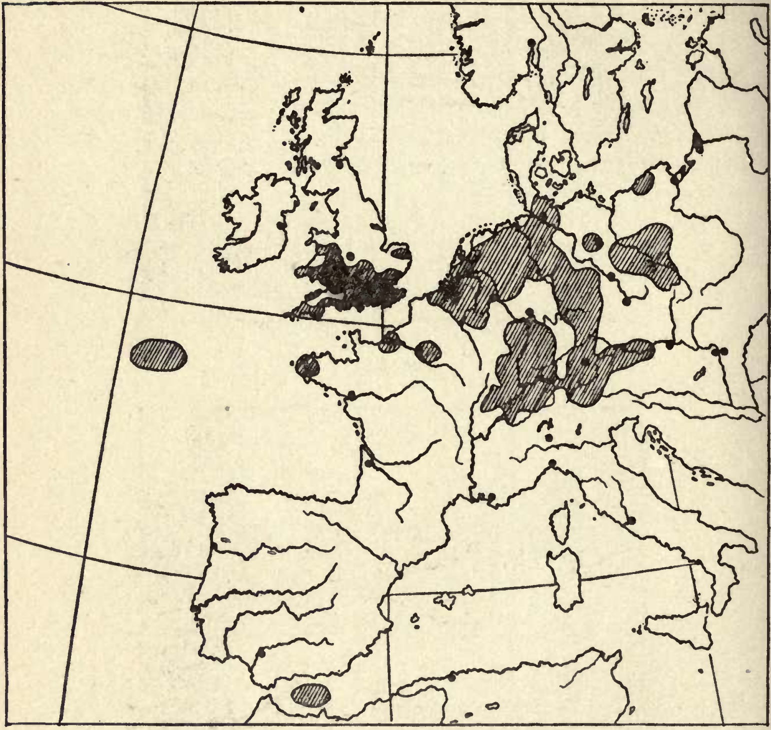

FIG. 35—DISTRIBUTION OF "BLOOD RAIN," FEBRUARY 18-22, 1903

African coast, as to whose origin there can be little uncertainty. Fig. 35 shows the area within which this peculiar dust fall was noted. No reports of dust were received from the northern part of the British Isles, where the trajectories show that the air was derived from the Western Atlantic. It is curious also that Spain and Southern France experienced no "blood rain," though the phenomenon was widely reported in Germany. These facts are all in agreement with the air circulation indicated by the trajectories constructed from the maps.

CHAPTER VIII

LINE SQUALLS

VERY interesting phenomena are met with along the line of separation between two air currents of different origins which are all the more marked if the two currents are of sensibly different temperature and if their directions are not approximately parallel, as in the case just considered, but inclined to one another at a considerable angle. An interesting example of this kind occurred on February 8th, 1906, and has been investigated in detail. Attention was first directed to the occurrence by a violent thunderstorm which occurred in London, shortly before 3 p.m. The storm was accompanied by a violent squall and a sharp fall of hail and snow, and the wind shifted abruptly from south-west to north-west. Pressure which had been falling steadily since the early part of the morning, increased in a few minutes by 3 millibars (·09 inch) and then remained approximately steady during the following hours while the temperature fell with equal suddenness from 44° to 37° and subsequently remained at the lower level. Reports of similar occurrences were received from other places and it was decided to collect all available information and to study the occurrence in detail. It appeared that thunderstorms accompanied by all the essential features just described had occurred throughout the eastern part of England, south of York. Careful examination of the records from places in the north and west showed that changes of a similar nature, though much less violent had occurred over the whole of the United Kingdom. In many cases there was only a smart shower, or a change of wind or a change from a falling to a steady barometer without a sudden increase of pressure, but every record examined gave some indication of an abrupt change in the meteorological conditions. When the times of occurrence of these changes were plotted on a map it was possible to connect them by a set of isochronous lines showing the spread of the phenomenon over the country. These lines are reproduced in Fig. 36.* The first indication, merely a sudden check to the fall of the barometer, was given at Stornoway in the Hebrides shortly after midnight, the last in the British Isles shortly before 6 p.m. (17·7h.) at Jersey. At a later date some records were also obtained from France which showed that the disturbance swept on in similar manner into the heart of that country. A barogram from the Puy de Dôme Observatory showed a sudden increase of pressure, even greater than that experienced in London, at 1 a.m. on the following day. It is clear that we have here to deal with a phenomenon sweeping forward on a linear front, in this case with an average velocity of about 38 miles per hour. As the violent squall of wind is one of the most characteristic, and often destructive, effects associated with such linear disturbances, the name *line squall* is generally assigned to them.

Let us now turn to the weather maps for the day. The morning map (Fig. 37) shows a depression centred off the Norwegian coast. The position of the squall line at this time is indicated by a dotted line crossing the centre of Ireland and south of Scotland. On the southern side of the dotted line the surface winds over

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* Figs. 36 to 38 from Vol. XXXII., *Quarterly Journal*, Royal Meteorological Society.

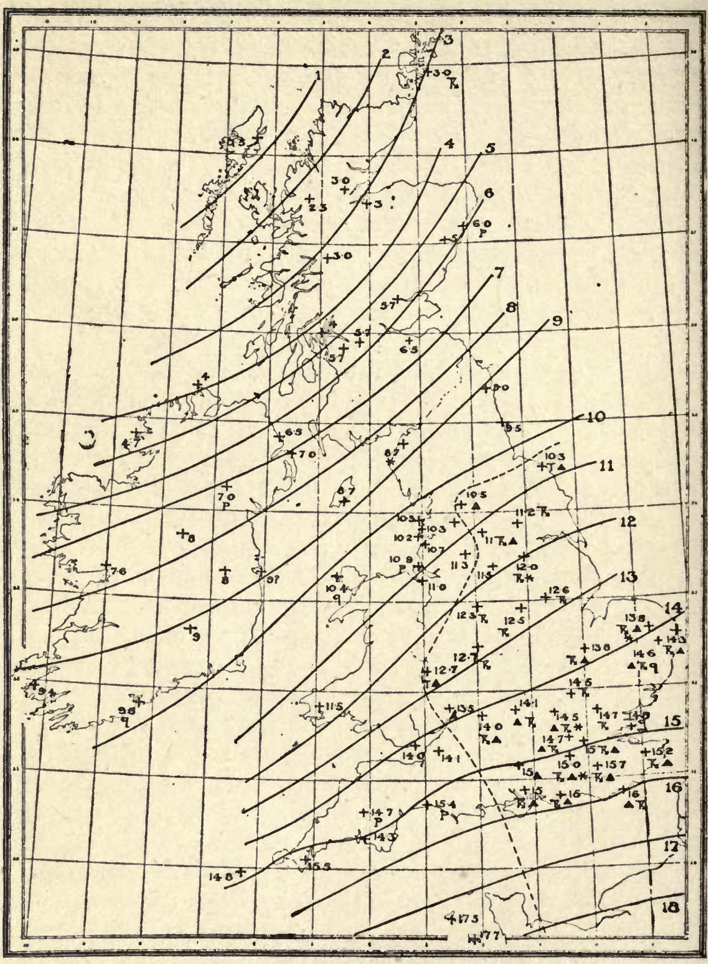


Fig. 36—Isochronous Lines for February 8th, 1906

England are all from south-west or west-south-west, while to the north of the line they are from west,

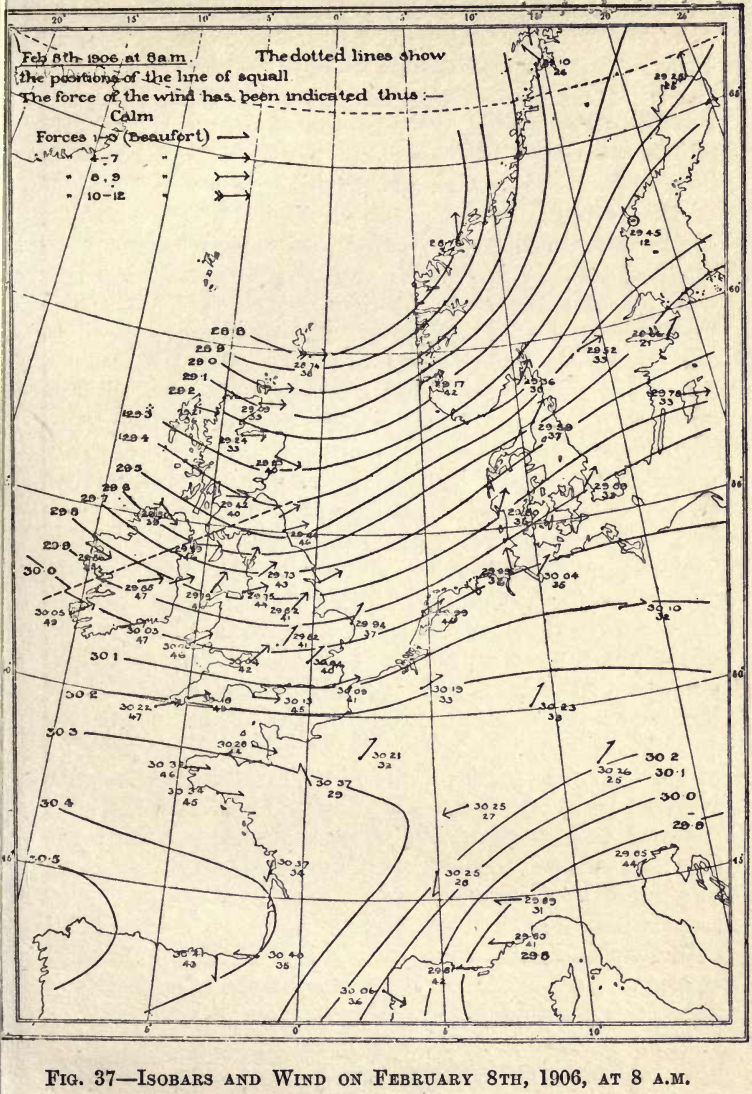


FIG. 37—ISOBARS AND WIND ON FEBRUARY 8TH, 1906, AT 8 A.M.

west-north-west or north-west. Fig. 38 shows the distribution over the British Isles at later stages, at 1 p.m., 2 p.m. and 3 p.m. The difference between the wind directions on either side of the squall line has become accentuated. The sudden increase of pressure which we noted in describing the London records is indicated on the map as a sudden dislocation or "fault" of the isobars along the squall line. The separation of the south-westerly and north-westerly winds is more marked than it was in the morning map and is borne out by the pressure gradients. In the south the gradient indicates air flow from west-south-west to east-north-east, while north of the squall line it shows a flow from north-west. We have here then an obvious case of a current from north-west impinging on a warmer current from west-south-west and forcing its way underneath it. The violent upward displacement of the warmer air of the west-south-west current gives rise to the heavy rain and hail, and incidental to this violent precipitation we have thunder phenomena. The maps for 2 p.m. and 3 p.m. reproduced in Fig. 38 show a second discontinuity of the isobars in the region of the north-westerly current in the rear of the principal change. At 2 p.m. this stretched from the south of Ireland to the Humber and by 3 p.m. it had progressed to a line from South Wales to the Wash. The passage of this secondary line some two hours later was again marked by a squall and by a repetition on a reduced scale of the weather phenomena associated with the first squall line. Such repetition of squall phenomena is a matter of common experience but the processes responsible for its occurrence are complex. They evidently indicate some instability in the north-west current. In London on this occasion the wind fell very light and backed temporarily to south-west shortly before the second squall, which was again accompanied by a sudden veer in direction to north-west. There

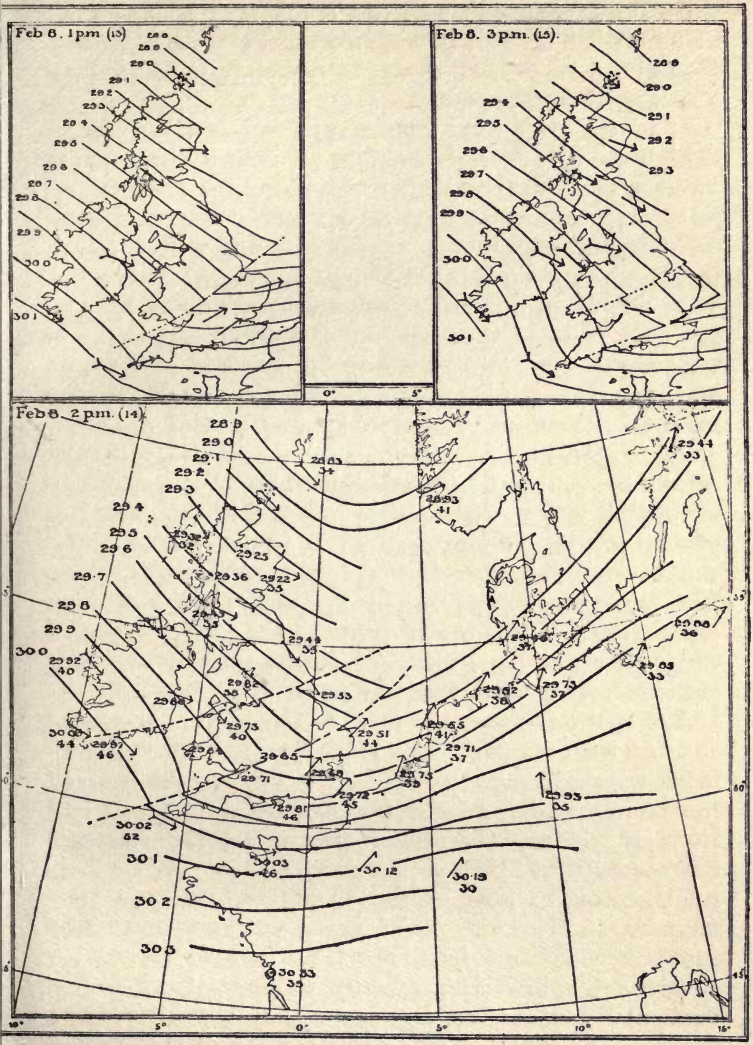
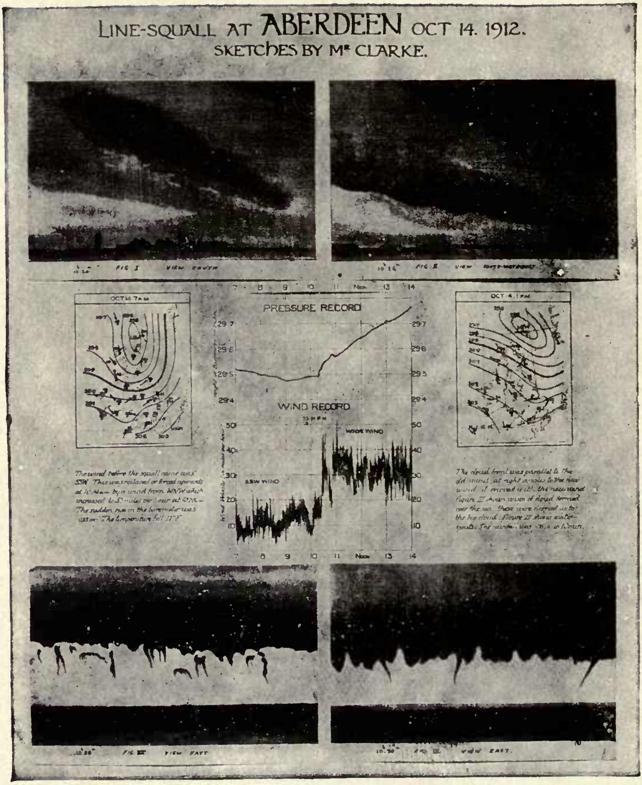


FIG. 38—ISOBARS AND WIND ON FEBRUARY 8TH, 1906, AT 1 P.M., 2 P.M., AND 3 P.M.

was even evidence of a third though very slight squall at about 8 p.m. and it was not until after its passage that the wind settled down to a steady current from about west-north-west.

It is not often that line squalls are as clear cut as in the case just described and as a rule it is not possible to trace them over so wide an area, but in most depressions regions of interference between currents can be identified which advance across the map with a linear front. For example in the depression of November 11–13, 1901, for which trajectories were shown in Fig. 30 the cutting in of the cold air brought by such trajectories as *F*, and *G*, when they become westerly winds on the south side of the depression, on the warm current from the south as presented by trajectories *K* and *L* gave rise to line squall phenomena on the southern side of the path. The marked increase in the intensity of rainfall which often occurs just before the period of steady rain in a depression comes to an end, to which the term "clearing shower" is sometimes applied, is a phenomenon of the line squall type and is a direct result of this interruption of the supply of warm air from south-east, south or south-west by a colder current from a more westerly point.

A **V**-shaped depression may be regarded as a special case of a line squall in which the two currents are from approximately opposite directions, the line of the trough which sweeps forward keeping parallel to itself represents the linear front. In the example represented by Fig. 8 the trajectories of the air on the eastern side of the trough represent a flow from south to north, while those on its western side indicate a current from north-north-west to south-south-east, which is gradually extending its régime eastward and forcing its way under the air coming up from the south. This displacement of the surface air by a colder current is not the only cause of pre-



The wind before the squall came was SSW. This was replaced or [illegible] upwards at 10.30 a.m. by a wind from WNW, which increased to 5[illegible] miles per hour at [illegible] — The sudden rise in the barometer was .033 in. The temperature fell [illegible]°F.

The cloud front was parallel to the old wind, at right angles to the new wind. It moved with the new wind. Figure III shows wisps of cloud formed over the sea; these were carried up to the big cloud. Figure IV shows water-spouts. The rainfall was [illegible] in 10 min.

FIG. 39

cipitation in a V-shaped depression. Rain generally sets in some hours before the passage of the trough due to convergence of air trajectories within the south wind, but in many V's the squall phenomena at the time of passage of the trough are very pronounced. In others they are less accentuated; the wind veers gradually from southerly to westerly or north-westerly directions, sometimes with a calm interval about the time of passage of the trough and the rain gradually ceases and ultimately the sky clears.

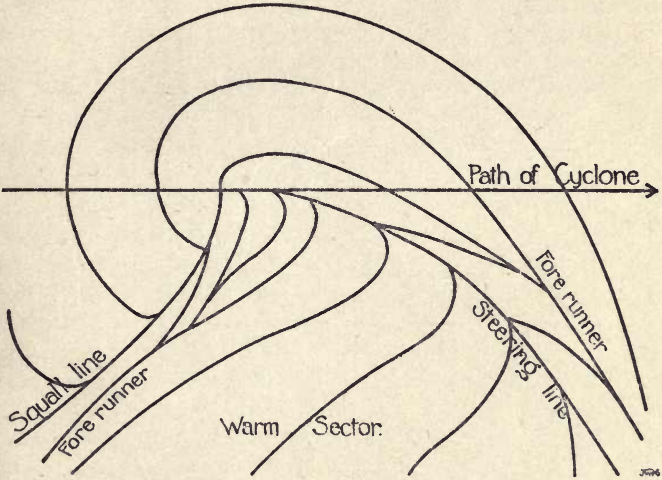
Fig. 39 reproduces some cloud sketches recording the passage of a line squall over Aberdeen Observatory at about 10.30 a.m. on October 14, 1912, in which the linear extension of the phenomenon was very conspicuously shown by the cloud formation. In this case the two wind currents were from south-south-west and west-north-west respectively, the latter wind being very much the stronger. The cloud front was parallel to the old wind direction (*i.e.*, from S.S.W. to N.N.E.) at right angles to the new wind and moved with the new wind.

Quite recently attention has been drawn to the phenonema arising when air currents from different sources meet, by the work of V. Bjerknes and his pupils. Deprived of weather telegrams from the greater part of Europe during the War, the Norwegian meteorologists found their isobaric charts, which only covered a very limited area, of comparatively little use for forecasting and this led to a serious attempt to use in practice a method of forecasting which Bjerknes had originally evolved from theoretical considerations. The method involved the collection of reports from a very large number of stations in the restricted area from which information could be made available. The observations of wind thus brought together were used for drawing the "lines of flow" of the wind, that is for showing the paths along which the air was apparently moving at the time when the observations were taken. This instantaneous picture of motion does not necessarily represent the actual trajectories of the air for reasons which have been considered at length in the previous chapter; it would only do so in the special case when the isobaric configuration remains unchanged, but it generally enables us to identify definite lines of convergence or divergence of air motion to which Bjerknes attributes great importance. A line of convergence is a line towards which the wind blows inwards from both sides and similarly a line of divergence is one from which the wind blows outwards on both sides.

According to Bjerknes two lines of convergence can be identified in a normal cyclone. Both lie on the right hand side of the path of the cyclone. One enters the cyclone from the front, *i.e.* from the south-east quadrant if the motion is towards the east, and can be traced right up to the centre of the system. The tangent to this line at its terminus in the cyclone centre seems to be identical with the path of the centre. As the line thus gives the direction of motion of the centre, if only the observations are sufficiently numerous for its complete identification, it has been called the *steering line.* The other line of convergence comes into the cyclone from the rear on the right hand side and is called the *squall line.*

Bjerknes thus arrives at the schematic representation of the flow of air in a cyclone shown in Fig. 40. The two lines of convergence mark off what may be called the "warm sector" from the remaining part of the system. Both lines move in the northern hemisphere to the right relatively to the direction of the wind along the line. The passage of the steering line thus gives rise to an increase of temperature, while the passage of the squall line is marked by a sudden decrease of temperature.

The squall line is a locus of undercutting of cold air along which heavy showers, rain and other line squall phenomena of the type described in the earlier sections of this chapter take place. The steering line on the other hand marks the intersection of the plane of separation between cold and warm air masses with the surface of the ground. We must imagine this surface



Lines of Flow in a Moving Cyclone after Bjerknes.

Fig. 40

of separation to slope upwards to the right of the steering line, so that there is an overflow of warm air over the cold air on the right of the steering line which gives rise to an area of steady rainfall in front of the steering line. The general picture of the air motion in a cyclone which is arrived at from these considerations is thus very similar to that derived from the trajectories of the depression of November 11–13, 1901, which were considered in detail in the preceding chapter (Fig. 30). The scheme is less accordant with that suggested by the looped trajectories of the more rapidly moving cyclone of March 24, 1902 (Figs. 28 and 29).

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CHAPTER IX

DIURNAL VARIATION OF WIND AND THE VARIATION OF WIND WITH HEIGHT

WE have seen already how the measurements of surface wind made with instruments exposed in the ordinary manner on buildings or even on specially erected high masts are influenced by the local conditions of exposure and the relief of the surrounding country, and we shall not be surprised to find a rapid increase of velocity with increasing elevation in the lower strata. Up aloft, at some level which we cannot define precisely, we may take it that the air is moving with velocity appropriate to the pressure gradient, which can be calculated from the equation of p. 61. In the layers near the ground the motion is much retarded by friction. At $\frac{1}{2}$ metre above the surface the velocity above grassland is only about half what it is at 10 metres, the standard height of exposure for an anemometer head. The air near the surface loses some of the velocity necessary to balance the pressure gradient and consequently yields to the gradient and its course is no longer along the isobars but inclined inwards towards the side of low pressure, as we saw in Chapter I. in considering the weather map and Buys Ballot's law.

The question is often asked whether there is any law according to which the approach from surface conditions to the undisturbed flow of the upper air is made. The problem has been investigated experimentally by comparing the readings of anemometers exposed at varying heights above ground, a method which obviously limits us to the investigation of the lowest strata or by discussion of the results obtained by raising anemometers by means of kites or by pilot balloons. Working by the first method with anemometers exposed over flat meadow land up to heights of about 100 feet Hellmann suggested a formula according to which the velocity varies as the fifth root of the height above ground while experiments with kites up to the 2,000 feet level conducted by Archibald so long ago as 1885 were best represented by a formula involving the fourth root of the height. Recent work with pilot balloons has emphasised the complexity of the phenomena involved and as the underlying principles have become more clearly recognised, have shown that we cannot hope to represent the results adequately by any simple formula. The whole question is intimately bound up with the eddying motion in the lower strata with which we have already been made acquainted when considering the gustiness as shown by an ordinary anemometer trace.

The theory of eddy-motion in the air has been worked out by G. I. Taylor and has given added insight into many of the facts of the motion of the lower strata of the air. In the course of the ordinary motion of the air, which we call wind, eddies of varying size are constantly being formed. While the spin of the eddy is maintained the mass of air involved is endowed by virtue of its spin with a special forced motion, but presently the eddy disintegrates and the air which has been forcibly transported in the eddy mixes with its surroundings. The effect of eddies is to promote mixing of adjacent layers of air, by which process the lowest layers gain in momentum at the expense of those above. If for any reason eddying is reduced the motion of the lower layers is checked by friction, while the higher layers are less disturbed, and we consequently find a fairly rapid increase of velocity with height. On the other hand as eddies develop, there is more interchange of air between adjacent layers, and the increase of velocity with height becomes smaller. The intensity of eddy motion depends on many factors. The configuration of the ground plays an important part which may be expected to vary with direction of the wind if the relief of the country is at all irregular. The temperature of the ground also plays its part. If the surface of the earth is cold, colder than the air passing over it, eddying is reduced, but if the ground becomes warmed above the temperature of the air, so that local convection is induced, the eddying is increased. The ratio of W the velocity at the level of the anemometer to G the gradient velocity, thus depends on many circumstances. It must be different over land from what it is at sea, and moreover must be different for different kinds of land surfaces; it must be different by day from what it is at night; it must vary with the season and according to the amount of cloud.

This variation in the intensity of eddy motion explains the well-known diurnal variation of the wind. The freshening of the wind on a fine day, as the sun increases in power and its dying down again as the sun sets is a matter of common observation. During disturbed weather general meteorological changes may obliterate the diurnal variation entirely, and on cloudy days it is so reduced as to escape notice but it can invariably be traced in the records of an anemometer if we work out the average velocities for each hour. The results for a selection of places are set out in the table on p. 104 for the months of January and July.

The highest and lowest values are set in special type. The figures given are averages computed from many observations; on individual days the contrast may be

DIURNAL VARIATION OF WIND VELOCITY.

MEAN VELOCITY OF THE WIND IN METRES PER SECOND FOR EACH HOUR IN JANUARY AND JULY

	1	2	3	4	5	6	7	8	9	10	11	Noon	13	14	15	16	17	18	19	20	21	22	23	24
Kew, January, (1881–1910) ..	3·3	3·3	3·3	3·3	3·4	3·4	3·3	3·4	3·5	3·8	4·2	4·3	4·3	4·3	4·1	3·8	3·8	3·7	3·7	3·7	3·6	3·5	3·4	3·4
Kew, July ..	1·9	1·8	1·8	1·8	1·8	2·2	2·6	3·0	3·4	3·7	3·9	4·0	4·1	4·2	4·1	4·1	3·9	3·6	3·2	2·7	2·4	2·2	2·1	2·0
Valencia, January (1881–1910)	6·5	6·4	6·4	6·3	6·4	6·3	6·4	6·4	6·5	6·4	6·3	6·9	7·2	7·2	7·2	6·9	6·7	6·5	6·5	6·4	6·4	6·6	6·6	6·6
Valencia, July ..	3·7	3·7	3·7	3·7	3·6	3·7	3·9	4·3	4·7	4·9	5·1	5·5	5·7	5·8	5·8	5·6	5·5	5·3	4·9	4·3	4·0	3·8	3·7	3·8
Helwan, January (1906–1910) ..	3·1	2·9	2·9	3·0	2·9	2·8	2·8	2·7	2·7	2·9	3·9	4·3	4·7	4·8	4·9	4·8	4·3	3·8	3·7	3·8	3·7	3·6	3·3	3·1
Helwan, July ..	3·7	3·4	3·6	3·3	3·0	2·9	3·2	3·7	4·1	4·0	4·5	5·0	5·5	5·7	6·1	6·3	6·6	6·7	7·0	7·3	7·3	6·7	5·7	5·0
Paris, January ..	2·1	2·1	2·2	2·2	2·2	2·2	2·2	2·3	2·3	2·5	2·7	2·8	2·9	3·0	2·7	2·6	2·3	2·3	2·2	2·3	2·3	2·2	2·2	2·2
Paris, July ..	1·4	1·3	1·3	1·4	1·5	1·6	1·9	2·1	2·3	2·6	2·8	2·9	3·0	3·0	2·9	2·8	2·7	2·4	2·1	1·7	1·7	1·6	1·6	1·5
Eiffel Tower, January ..	10·8	10·8	10·8	10·8	10·8	10·9	10·9	10·9	10·7	10·3	9·9	9·5	9·1	8·9	9·2	9·4	10·0	10·6	10·7	10·9	11·0	10·9	11·0	11·0
Eiffel Tower, July	9·0	9·1	8·7	8·7	8·6	8·1	7·2	6·3	6·1	6·5	6·8	7·2	7·3	7·7	7·3	7·6	7·7	7·7	7·8	8·0	8·8	9·1	9·3	9·2

much greater or on the other hand it may be almost entirely eliminated. Confining our attention to the anemometers near the ground we find, as we should expect from what has been said above about the influence of the warming of the ground on eddy motion, that the minimum velocity occurs at night and the maximum about midday or in the early afternoon, and also that the effect is much greater in summer than in winter. At Kew in January the average velocity during the night is only about 3·3 or 3·4 metres per second, but soon after sunrise it begins to increase and is at maximum value, 4·3 metres per second from noon until 2 p.m. In July the contrasts are rather greater. The minimum velocity, 1·8 metres per second lasts from 2 a.m. till 5 a.m., but by 2 p.m. in the afternoon the average velocity has more than doubled. At Valencia, as we should expect in so exposed a part as the west coast of Ireland, the wind is generally considerably stronger than at Kew, but as we should also expect at a coast station where the variations of wind partake to some extent of the nature of those met with at sea, the diurnal variation is relatively small. In January the minimum and maximum values are 6·3 and 7·2 metres per second respectively, an increase of only about one sixth. In July the increase is somewhat greater, from 3·6 to 5·8 metres per second. It is interesting to compare these values taken from records in our own country with those for Helwan near Cairo, which are also given in the table. We have there to deal with a climate where the solar influence is very strong and we should expect strong eddy-motion to develop during the day. It is curious to note how the time of maximum wind velocity is delayed in July until 8 or 9 p.m., and how after the maximum has occurred there is a gradual decrease of turbulence throughout the night. The minimum velocity is not reached until 6 a.m.

The last line in the table shows the average diurnal variation of the wind as recorded by the anemometer on the Eiffel Tower at a height of 305 metres, roughly 1,000 feet above the ground. It will be seen that at this level the phenomena of diurnal variation are quite different from what they are at the surface. Instead of occurring in the early afternoon, the maximum velocity occurs about midnight both in winter and summer, while the minimum velocity which at the surface occurs in the night, falls to the hour 9 a.m. in summer and 2 p.m. in winter. We must attribute the minimum by day to the mixing of air of which the motion has been checked by surface friction, with the air at the 1,000 feet level in consequence of the enhanced eddy motion which occurs by day. At night when eddy motion dies down there is less admixture of air from near the surface with the air at the 1,000 feet level; under some conditions there may be none at all in which case we may regard the velocity registered by the anemometer of the Eiffel Tower as a measure of the undisturbed flow of air which it should be possible to calculate from the pressure gradient by the equation of p. 61. In the day time the fact that the velocity reaches a minimum shows that this cannot be the case, the disturbing effects of the surface must then extend to heights greater than 1,000 feet.

Consideration of the results from the Eiffel Tower will lead us to expect that the rate of variation of wind velocity with height must show a considerable diurnal variation. Generally speaking we should look for a rapid increase of velocity with height at night or in the early morning, when eddy action is relatively small and a more gradual increase, but associated with stronger winds near the surface, about midday when conditions favour more vigorous eddy action. This expectation is borne out by observation and in illustration thereof we may quote some results of pilot balloons obtained by J. S. Dines at South Farnborough. If the ascents be arranged in groups according as they were made in the early morning, about 7 a.m., midday, about 1 p.m., or during the afternoon, about 4 p.m., they give the following mean velocities at different heights above the ground.

Height above Ground in Metres.	7 *a.m.* 14 *ascen's.*	1 *p.m.* 17 *ascents.*	4 *p.m.* 16 *ascents.*
50	4·7	6·1	5·2
100	6·5	7·2	6·6
150	8·0	8·0	7·7
200	8·9	8·5	8·6
250	9·7	8·9	9·2
300	10·1	9·2	9·9
400	11·1	10·1	10·9
500	12·2	10·8	11·3

It will be seen that whereas the early morning ascents show an increase from 4·7 to 12·2 metres per second in the height range from 50 to 500 metres, the midday ascents only range from 6·1 to 10·8 metres per second while the afternoon ascents are intermediate between the two, ranging from 5·2 to 11·3 metres per second.

Clearly where so much depends on factors which can only be partially known we cannot give a definite limit for the height to which the influence of the surface extends. It must depend on the strength of the wind as well as on local circumstances. As an example we may quote some results obtained from pilot balloon ascents by G. M. Dobson in a very open situation on Salisbury Plain, 600 feet above sea level, where conditions were not complicated by obvious cliff eddies or the uncertainties introduced by the transition from land to sea such as may be expected at stations on the coast. The occasions were classified as those of light winds, moderate winds or strong winds according as the velocities observed at 650 metres were below 4·5 metres per second, between 4·5 and 12 metres per second or above the latter limit.

The light winds which we may regard as comparatively free from eddy turbulence showed on the average little change of velocity with height. The moderate winds showed rapid increase at first and approximated to the gradient value at about 300 metres (under 1,000 feet) above the surface, while the strong winds in which eddy turbulence would be still greater did not approximate to the gradient value until close on 500 metres (1,500 feet) was reached. The approximation to the gradient direction was not attained until considerably greater altitudes were reached, somewhere about 800 metres above the surface. This difference in the behaviour of the two elements direction and velocity in the matter of the level at which approximation to the value appropriate to the gradient is reached is in accordance with Taylor's theory of eddy motion. The theory also leads to a relation between the ratio of the velocity at the surface to the gradient velocity and the angle a between the direction of the surface wind and the undisturbed gradient wind

$$\frac{W}{G} = \cos a - \sin a.$$

When applied to the Salisbury Plain observations the results deduced from this formula were as follows:—

	Light Winds.	*Moderate Winds.*	*Strong Winds.*
Observed value of $\frac{W}{G}$ ..	·72	·65	·61
Observed a	13°	21½°	20°
Calculated a	14°	18°	20°

The closeness of the agreement is remarkable when we bear in mind the inevitable uncertainties of the measurements.

It should be noted that these numerical results represent average values deduced from many observations. If we confine ourselves to considering individual occasions we must not be surprised if we often find considerable differences between the gradient wind and the results of observation as disclosed by pilot balloons. The gradient velocity or rather the geostrophic velocity (see p. 60) is often quoted in weather reports to aviators and others as " the computed velocity at 1,500 feet," and for brevity the word *computed* is often omitted. It represents the best estimate which can be given in the absence of actual measurement at some such level. It should, however, be borne in mind when making comparison with pilot balloon results that the gradient which we can measure from the isobars on our maps is the gradient at sea level, and that it may require appreciable correction to render it applicable even at so low a level as 1,500 feet. If information for higher levels is required an even greater correction may be needed. In cases when the isobars are markedly curved the cyclostrophic part of the gradient (see p. 60) may be appreciable but unless we have maps for very short intervals from which the trajectories of the air may be worked out, we have no data from which the cyclostrophic correction can be computed. On the other hand the velocity for the 500 metre level computed from a pilot balloon ascent can only be accepted as an index of the general velocity at that level for the neighbourhood of the ascent with some reservation. If the ascent be worked out from the readings of only one theodolite, as is generally the case, vertical movements of the air due to cliff eddies, convection near clouds, and other causes cannot be detected but they may introduce appreciable errors in the computed horizontal velocities. Leakage of gas from the balloon, not to mention actual errors of observation also leave an appreciable margin for uncertainty. In quoting the results of a pilot balloon ascent for the guidance of a pilot we must, therefore, reckon with the possibility that the ascent is not a representative one for the course over which it is proposed to fly.

CHAPTER X

THE WINDS OF THE FREE ATMOSPHERE

ABOVE the first kilometre or two the influence of the surface turbulence may be regarded as eliminated but the structure of the atmosphere remains very complex so that no satisfactory simple generalisation of the results of observation by the method of taking mean values can be given. It should be remembered that the observational material available for considering the problem is itself to some extent selected. If we use kites, we are restricted to occasions when there is sufficient wind to lift the kite and not so much that it breaks the kite wire. For information from the highest layers we depend mostly on the results of pilot balloon ascents or observations of the drift of shrapnel smoke and these can only be observed at great heights in the absence of low cloud. If cloud be absent occasions of strong wind are unfavourable for high ascents with pilot balloons as the balloon gets carried beyond the range of vision, while still at fairly low altitudes. Even observations made by watching the drift of smoke from shell bursts at prearranged altitudes, though they can often be fitted into clear patches of sky are liable to be spoiled by cloud.

The data available for the region free from surface turbulence show that we may distinguish cases in which there is little further change either in direction or velocity up to great heights once the undisturbed flow has been reached. Other cases show steadily increasing velocity with height without much change of direction, or again we may find a gradual change both in direction and velocity. Above the level of ten kilometres a decrease of velocity is very frequently met with, especially in the case of strong winds. This question will be considered more fully in Chapter XIV. We may here draw attention to a tendency for the velocity to decrease with height which is exhibited by many winds from east; not infrequently this decrease begins to take place even before the normal increase in the layers nearest the ground has brought the velocity up to the value appropriate to the gradient. Easterly currents are often of little vertical thickness and it is by no means uncommon to meet with a wind from the west above a surface current from the east. Instances of this must have often come under the notice of every regular observer of the drift of clouds.

Material for the study of the upper winds is being collected rapidly in obedience to the insistent demands of aviation. Much information for the British Isles, most of it obtained with pilot balloons, is set out day by day in the Upper Air Supplement to the *Daily Weather Report* which gives a series of maps setting out the data available for different levels up to 15,000 feet from ascents made about 7 a.m., 1 p.m. and 6 p.m. each day.

The results obtained with balloons and kites make it obvious that we cannot expect to get any useful information regarding the wind at great heights, say several kilometres above the surface by any process of extrapolation from observations made with anemometers near the surface. The pressure distribution as revealed by the isobars enables us to calculate the gradient velocity which may be taken as a reasonable representation of the air flow at moderate heights such as 500 metres or perhaps a kilometre or two, but for higher levels the pressure distribution at the surface affords but little guidance.

If we had the means of determining the distribution of pressure at the high level we might use it for determining the direction and velocity of the wind by calculation for we know that the motion of the air at any level provided it is " steady " is related to the pressure gradient at that level by the gradient equation set out on p. 59. Unfortunately we have not the means of evaluating the gradient at high levels from observations made at the surface. It is true we may use the formula given on p. 33 in connexion with the calculation of height from observations of pressure to find an approximate value for the pressure at any level but in order to obtain sufficiently accurate values for the construction of a pressure map for, say, the level 4,000 metres above sea level we must have satisfactory information regarding the variation of temperature with height from many stations. If a sufficient number of reporting stations could supply, in addition to the usual surface observations, the results of a registering balloon giving the variation of temperature with height we might calculate the pressure at 4,000 metres for a sufficient number of points on the map to enable us to represent the pressure distribution at that level and hence to calculate the gradient wind appropriate to it.

The construction of such maps for high levels is as yet not practicable but we may use an observation of wind in the upper air obtained by pilot balloon or shell burst to calculate an approximation to the gradient of pressure prevailing at the corresponding level in the atmosphere, and may then speculate as to the distribution of temperature required to produce it.

The gradient equation

$$v=\frac{\gamma}{2\omega D\sin\phi}$$

involves D the density of the air, which is connected with the temperature and pressure by the well known equation expressing Boyle's and Charles's laws

$$D=\frac{1}{R}\cdot\frac{p}{\theta},$$

where R is a numerical constant, p and θ the pressure and temperature, the latter being expressed on the absolute scale. Substituting for D and disregarding the cyclostrophic term we may write the gradient equation

$$v=\frac{R}{2\omega\sin\phi}\cdot\frac{\theta}{p}\cdot\gamma=K\frac{\theta}{p}\gamma.$$

Our object is to evaluate γ, the gradient at the level to which the observed velocity v applies. The term $\frac{R}{2\omega\sin\phi}$, for which we may write a constant K, can be evaluated as we can assign values for R, ω and ϕ, but at first sight it appears that further progress in the determination of γ is blocked as we do not know θ and p the temperature and pressure at the level for which v is observed. We observe, however, that we only need to know the ratio $\frac{\theta}{p}$ for evaluating γ and we shall see in a later chapter that above the level of three kilometres there is a close connection between the pressure and temperature of the air.

Observations have shown that if on any given occasion the pressure at any level is above the average for that level then the temperature will also be above the average temperature for that level, and similarly if pressure is below the average, temperature is also below the average. The result is that the ratio $\frac{p}{\theta}$ for any level above three kilometres shows remarkably small fluctuations from day to day and, therefore, we may use the average value of the ratio in calculations for individual cases, such as those we are now discussing without introducing serious error. The following table gives the mean values of the ratio $\frac{p}{\theta}$ for different levels ("Computer's Handbook"), where p is expressed in millibars θ in degrees on the absolute scale.

Height in Kilometres.	$\frac{p}{\theta}$	*Height in Kilometres.*	$\frac{p}{\theta}$
3	2·63	12	0·90
4	2·36	13	0·77
5	2·12	14	0·65
6	1·90	15	0·54
7	1·70	16	0·45
8	1·52	17	0·38
9	1·35	18	0·32
10	1·19	19	0·28
11	1·04	20	0·28

By substituting in the equation given above the value of $\frac{p}{\theta}$ appropriate to the level for which we have an observation of the velocity, we can calculate an approximate value for the pressure gradient at that level; in other words for the distance between consecutive isobars drawn for a specified interval.

We may illustrate the process by reference to an example for an occasion which has become notorious, October 19th, 1917, when a fleet of Zeppelins was carried out of its course and lost over France after an air raid on London, by an unexpected strong upper current from north. The distribution of pressure at the surface has been shown in Fig. 10, p. 22, as an illustration of an anticyclonic wedge. Over England and also over Eastern Scotland the winds were light, generally between 5 and 10 miles per hour, the general direction being from north-west. The distance between the isobars at the surface for 1,025 and 1,020 millibars, over South-eastern England was 480 kilometres, corresponding with a gradient wind of 16 miles per hour a value which is in reasonably good agreement with the velocities for the 1,000 feet level observed by pilot balloons, which were of the order 15 miles per hour.

For the next few thousand feet of height the pilot balloons show no conspicuous change, but at 10,000 feet, above 3,000 metres, the wind régime changed. The wind shifted towards north and increased in strength very materially. Pilot balloon ascents at these high levels are available for Portsmouth and from near Edinburgh and gave the following observations of wind, the velocities being expressed in metres per second to facilitate computation.

At 3000	metres	Portsmouth	N.N.W. 11·2	m/sec.	Edinburgh	N. 9·4	m/sec.
,, 4500	,,	,,	N. by W. 27	,,	,,	N. 18	,,
,, 6000	,,	,,	N. by W. 40	,,	,,	—	,,

The value of the constant $\frac{R}{2\,\omega \sin \phi}$ appropriate to the latitude of Portsmouth is 25 and at 6,000 metres the table just quoted gives 1·90 as the appropriate value for the ratio $\frac{p}{\theta}$.

The pressure gradient over Portsmouth at the 6,000 metre level was thus:

$$\frac{1}{25}\times 1{\cdot}90\times 40=3{\cdot}04 \text{ millibars in 100 kilometres,}$$

or the distance between consecutive isobars drawn for intervals of 5 millibars would be 164 kilometres or about 100 statute miles, and the run of the isobars would be from N. by W. to S. by E.

If we wish to draw maps of isobars for the level 6,000 metres we have at any rate this for guidance that the distance between isobars drawn for intervals of 5 millibars must be about 100 miles in the Portsmouth area.

The pressure gradients in millibars per 100 kilometres corresponding with the remaining wind observations quoted above are as follows:—

At 3000	metres	over	Portsmouth	1·16,	over	Edinburgh	1·04
,, 4500	,,	,,	,,	2·39,	,,	,,	1·66
,, 6000	,,	,,	,,	3·04,	,,	,,	—

or if we prefer to specify gradients in the distance between successive isobars drawn for the interval 5 millibars

At 3000	metres	over	Portsmouth	430 km.,	over	Edinburgh	480 km.
,, 4500	,,	,,	,,	209 ,,	,,	,,	300 ,,
,, 6000	,,	,,	,,	164 ,,	,,	,,	—

We see therefore that we can use an observation of the direction and velocity of the upper wind to evaluate the pressure gradient at the level at which the observation is made and the results of a pilot balloon may therefore be expressed in terms of the variation of pressure gradient in the vertical. We may then go a step further and reason as to the distribution of temperature necessary to bring about the variation of gradient. Let p be the pressure (in millibars) θ the temperature (on the absolute scale) and D the density (in grammes per cubic centimetre) at a point A in the atmosphere and $p+\Delta p$, $\theta+\Delta\theta$, $\mathrm{D}+\Delta\mathrm{D}$ the corresponding values at a neighbouring point B at the same level, and let us suppose that the line joining A and B is taken along the pressure gradient (*i.e.* at right angles to the isobars). Consider the variation of pressure in the vertical. If δp be the change of pressure over A corresponding with a height step δh we have

$$\delta p = -gD\delta h,$$

and for the position B

$$\delta(p+\Delta p) = -g(\mathrm{D}+\Delta\mathrm{D})\delta h.$$

The change in pressure difference between A and B corresponding with the height step δh, *i.e.* the change in pressure gradient which can be evaluated from the data supplied by a pilot balloon ascent, is therefore

$$-g\,.\,\Delta\mathrm{D}\,.\,\delta h.$$

It depends on $\Delta\mathrm{D}$ the difference in density of the air at A and B respectively. The density, as we know, is determined by the pressure and temperature, in accordance with the relation, already made use of, expressing the fundamental gaseous laws

$$\frac{p}{\theta}=\mathrm{R}\cdot\mathrm{D}.$$

We have therefore to select suitable values for the differences of pressure Δp and temperature $\Delta\theta$ between A and B to give the required differences in density. Δp is the pressure gradient which we can calculate from the observed wind and hence the only remaining unknown quantity is $\Delta\theta$, the horizontal gradient for temperature. It can be shown that the rate of increase of pressure difference in millibars per metre of height is given by the expression*

$$\cdot 0342\,\frac{p}{\theta}\left(\frac{\Delta\theta}{\theta}-\frac{\Delta p}{p}\right).$$

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* From the fundamental gas equation $\frac{p}{\theta}=\mathrm{R.D.}$ where R is numerically equal to $2\cdot869\times10^6$C.G.S. units for dry air.

Hence $$\frac{\Delta p}{p}-\frac{\Delta\theta}{\theta}=\frac{\Delta\mathrm{D}}{\mathrm{D}},$$

or $$\Delta\mathrm{D}=\frac{p}{\mathrm{R}\theta}\left(\frac{\Delta p}{p}-\frac{\Delta\theta}{\theta}\right).$$

But $$\delta(\Delta p)=-g\times\Delta\mathrm{D}\times\delta h,$$

whence $$\frac{\delta(\Delta p)}{\delta h}=\frac{g}{\mathrm{R}}\,\frac{p}{\theta}\left(\frac{\Delta\theta}{\theta}-\frac{\Delta p}{p}\right),$$

or substituting numerical values for g and R

$$\cdot 0342\,\frac{p}{\theta}\left(\frac{\Delta\theta}{\theta}-\frac{\Delta p}{p}\right).$$

As we have seen above we may regard the ratio $\frac{p}{\theta}$ as approximately known; numerical values were given above. The values of θ and p appropriate to the individual case can only be obtained from the results of an ascent of a registering balloon carrying barograph and thermograph. In the absence of such direct observational results we can use the average values appropriate to season and level to obtain some idea of the magnitude involved. If we proceed in this manner with the data for October 19th, 1917, quoted above we find that in order to produce the change in pressure gradient over the Portsmouth region from 1·16 millibars per 100 kilometres at 3,000 metres to 2·39 millibars per 100 kilometres at 4,500 metres, the appropriate value for $\Delta\theta$ must be about 3 degrees per hundred kilometres, in other words, at the level of about 3,000 to 4,000 metres temperature decreased by about 3 degrees (absolute) per 100 kilometres towards the east. The further increase in the gradient from 2·39 to 3·04 millibars per 100 kilometres in the height interval from 4,500 to 6,000 metres again demands a somewhat similar gradient for temperature from west to east in the layers concerned. The exact figures are of little importance as too many assumptions have been made in arriving at them but the interest lies in the inference that the strong north wind up aloft must have been associated with a sharp decrease of temperature in the upper air over the North Sea as compared with the values over Central England at the same level.

In the Northern Hemisphere increase of pressure gradient with increasing height is associated with a horizontal gradient for temperature at right angles to the flow of air, the temperature being lower on the left hand side of the path. Some difficulty arises if there is change of direction of the pressure gradient with change of altitude as well as change of magnitude. In such cases we may proceed by resolving the gradient and the associated air motion into two components, the one representing motion in the north-south direction, the other motion in the east-west direction. The components may then be considered separately and the resultant conditions of motion be arrived at by compounding the effects on the two components.

In general the effect of low temperature in any region may be regarded as adding to the lower wind, as higher levels are reached, a superposed wind which flows round the centre of low temperature in a cyclonic direction.

CHAPTER XI

THE PHYSICAL PROCESSES OF THE ATMOSPHERE

WHEN discussing the changes of level of air in connexion with its horizontal trajectories it was pointed out that air expands when rising to a higher level in consequence of the reduction of pressure which it experiences and that cooling is a necessary consequence of this expansion. Conversely, if air passes from a higher to a lower level, it becomes warmer by reason of the compression which it undergoes in the process. If heat is neither communicated to the air from without or given up by it to its surroundings the warming or cooling is at the rate of one degree centigrade for a change of level of 100 metres in the case of "dry" air, using the word "dry," not in the chemical sense of free from all trace of water but to indicate that the air is unsaturated with moisture, in other words that it is in the state in which it conforms to the well-known gaseous laws of Boyle and Charles when subjected to changes of pressure and temperature.

Changes in the pressure, temperature and density of a sample of air, or other gas, taking place under the condition that there is no transference of heat to or from the gas are said to take place under *adiabatic* conditions. The conditions may be regarded as practically satisfied in the case of changes of level of air occurring in the free atmosphere. If we could conceive an atmosphere in which the distribution of temperature in the vertical had been attained entirely as a consequence of the vertical motion of air, we should expect to find the rate of decrease of temperature with height in it that corresponding with the adiabatic condition, viz. 1·0° per 100 metres.

The rate of decrease of temperature with height has often been referred to as the vertical temperature gradient, but it is preferable to restrict the word gradient to mean the rate of variation of an element, such as pressure or temperature, horizontally. The word *lapse* has been suggested to denote the decrease of temperature or pressure with height, and in what follows we shall adhere to this convention.

Let us consider what would happen in an atmosphere in which the lapse rate is the adiabatic if for any reason a part of the air got warmed. Such warming would reduce the density of the air, convection would come into play and the warmer air would commence to rise. In rising it would be cooled at the adiabatic rate and consequently at each level it would find itself still slightly warmer and therefore lighter than the surrounding air and it would continue to rise until it reached the top of that section of the atmosphere in which the adiabatic lapse rate prevailed. Conversely if any part of the air were cooled, the resulting increase of density would cause it to descend and again the downward motion would continue until the cooled air reached the surface or the lower limit of the part of the atmosphere in which the lapse rate was adiabatic. If on the other hand the lapse rate is less than the adiabatic, no such continuous ascent or descent would result from warming or cooling. In the case of warmed air, the initial decrease of density might be sufficient to start upward motion, but the cooling due to expansion with increasing altitude would be more rapid than the lapse rate in the surrounding air. Thus the ascending air sooner or later would find itself in equilibrium with its surroundings and the vertical motion would be arrested. Similarly if a portion of the air were cooled, the resulting increase in density might induce downward motion, but in this downward motion the air would become warmed again at the adiabatic rate, and presently the motion would again be arrested. The adiabatic is the maximum lapse rate which can occur in the free atmosphere. If we imagine an atmosphere in which a lapse rate greater than the adiabatic has got temporarily established, the conditions would be unstable. Convection would immediately come into play and a more or less violent readjustment, such as we have evidence of in thunderstorms, would restore a more stable arrangement.

The adiabatic lapse rate of 1 degree per 100 metres is a limiting value for unsaturated air. But the air is never dry in the chemical sense. A certain amount of water vapour is always present and the cooling due to ascent of air may be so great that the temperature of the air becomes reduced to the dew point. If the upward motion continues, further cooling takes place and results in the condensation of some of the water vapour on dust particles or other nuclei floating in the air in the form of cloud. If carried further the process leads to rain. If the dew point is passed and condensation occurs as a result of further cooling the latent heat of condensation reduces the rate at which temperature decreases with height. For this reason the adiabatic lapse rate in saturated air at temperatures such as ordinarily prevail near the ground is only about half what it is in unsaturated air, but the rate is dependent on the temperature. At low temperatures when the total amount of water which the air can hold is small, the lapse rate for saturated approximates to that for unsaturated air. This will be clear from an inspection of Fig. 41, in which the dotted lines show the adiabatic lapse rate for saturated air. The short full lines between the ground and the level of 1,000 metres show the adiabatic lapse rate for unsaturated air.

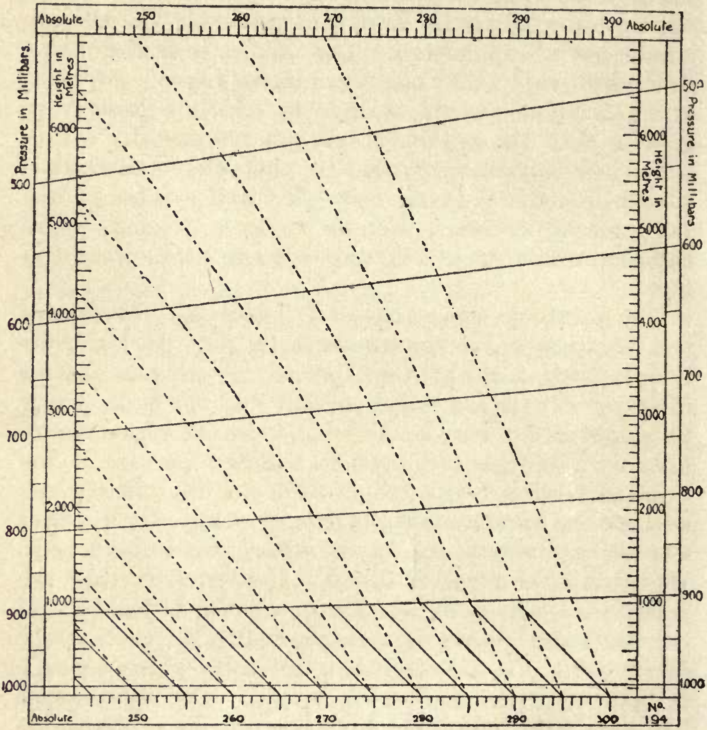


FIG. 41

Diagram showing the pressure in the upper air corresponding with the standard pressure (1013·2 mb.) at the surface and adiabatic lines for saturated air referred to height and temperature. (From Neuhoff, Smithsonian Miscellaneous Collections, Vol. 51, No. 4, 1910.)

So long as we are dealing with unsaturated air undergoing changes of level under adiabatic conditions the temperature of a given sample of air is independent of its past history and depends only on the level at which the air finds itself. If we start with air at 295a (295 on the absolute scale) at surface level and cause the air to ascend to any level, short of that at which saturation is reached, the temperature will still be 295 when the air is brought back to the surface if the adiabatic condition has been satisfied throughout. But if once the air is cooled so much that the saturation limit is attained and passed so that the water may be condensed and lost as rain this is no longer the case. An example will make our meaning clear. Suppose we start with air at the surface at a temperature 295a (72° F.) and of relative humidity 50 per cent. It will then contain 10 grammes of water vapour in each cubic metre of air. If this air be lifted the temperature will fall at the adiabatic rate of 1 degree per 100 metres until the level 1,000 metres is reached. The temperature will then be 285a, which is approximately the dew point of the air. The relative humidity will then be 100%. If the air be lifted still further the temperature decreases at the reduced rate of about half a degree per 100 metres, as indicated by the appropriate dotted line of Fig. 41, and throughout this stage the air will be saturated. When the level 2,000 metres is reached the temperature will have decreased to about 280a, and each cubic metre will now contain only 7·5 grammes of water instead of the original 10 grammes. The difference; 2·5 grammes will have been condensed from each cubic metre, in the form of cloud. The water drops of which the cloud consists will gradually settle under the influence of gravity and if sufficient time be allowed to elapse, will fall clear of the sample of air we are considering. If the air be then brought down again to the surface the temperature increases throughout the downward path at the full adiabatic rate for dry air, 1 degree per 100 metres, and thus the temperature when it gets back to the surface will have increased to 300a (81° F.), while each cubic metre will have lost 2·5 grammes of water. The relative humidity at the end of the cycle of operations will be only about 40 per cent. as compared with the original value 50 per cent. If the return journey to the surface had been commenced before all the condensed water had settled the first effect of the warming due to descent would have been to cause the cloud to evaporate and so long as there were water drops available, the warming would not have been at the full adiabatic rate for dry air so that the final temperature when the surface was regained would have been slightly below 300a.

The cooling inevitably associated with the ascent of air from lower to higher levels is the principal cause of the earth's rainfall. The most obvious example of its action is furnished by so-called orographical rains to which reference was made on p. 82. We have come across almost equally obvious cases when considering the trajectories of air in cyclonic depressions when we found evidence of currents of warm moisture laden air, raised from the surface over the top of a mass of colder air. There is no other physical process which can give rise to sufficient condensation to produce appreciable precipitation. The mixing of cold with warm air suggests itself as a possible source of rainfall but the amount of precipitation produced in this way is too small to be of importance when considered as rainfall. A cubic metre of air at 275a, if saturated contains 5·54 grammes of water; at 295a the amount would be 19·22 grammes. If we could by some process induce mixing in equal proportions of saturated air at these temperatures the mixture should contain 12·38 grammes per cubic metre. Now air at 285a is saturated when it contains 10·57 grammes per cubic metre. We might therefore expect that the difference 1·81 grammes of water would be condensed from each cubic metre. Actually less would be condensed for the final temperature of the mixture would be above 285a owing to the setting free of the latent heat of the condensed water. Even if the process of mixing could be carried on through a height of 500 metres the yield on our assumption would be less than 500 × 1·81 or 905 grammes of water for each square metre of area, equivalent to a rainfall of 0·9 millimetre. The assumptions which we have made are far more favourable to the formation of rain than anything which actually occurs in nature. Two supplies of air differing in temperature by 20 degrees on the absolute scale and of vertical height 500 metres are never found in sufficient proximity to one another to mix. Moreover, as we have seen in considering the trajectories of air motion, if currents of different temperature do impinge on one another, they do not mix. What actually occurs is that the warmer, and therefore less dense current rises up above the colder one. We may therefore dismiss the mixing of air supplies at different temperatures as a source of rainfall though as we shall see later it may result in the formation of fog of appreciable thickness. The amount of rainfall obtainable from an ascending current is practically unlimited, for the process of condensation continues as long as the current continues. It should of course be understood that by an ascending current we do not mean a current directed vertically upwards but merely that the current has an upward component. The actual motion may be and generally is up a very gradual incline.

If the upward component be large it may prevent the drops formed as a result of the cooling produced by the ascent from falling to the ground. Once formed a drop is subject to the action of gravity and in still, eddy-free air would ultimately fall to the ground if it did not evaporate before reaching it. The rate of fall depends on the size of the drop and is very small for small drops, and a comparatively small vertical component of the air motion will suffice to keep the drops suspended. The rate of fall increases with the size of the drops. Lenard has shown experimentally that there is an upper limit to the size of the drops which we can expect to meet with in rain. He allowed drops of water to fall into a current of air produced by a rotating fan and directed vertically upwards. By carefully adjusting the intensity of the blast he succeeded in keeping drops suspended for some seconds in the ascending current. Under such circumstances the rate of fall of the drop must be equal to the velocity of the ascending air. The latter could be measured by a suitable anemometer and the size of the drops was determined by catching individual drops on blotting paper and measuring them. He found that drops of about 5 millimetres in diameter remained suspended in an upward current of 8 metres per second, and we may therefore take it that this represents the rate of fall of a drop 5 millimetres in diameter. Larger drops which fall more rapidly, were invariably found to break up into a number of small drops. We may infer that similar processes go on in the atmosphere. Water drops, when they have grown to a size of about 5 millimetres, break up into smaller drops. It follows that no rain can fall through a wind having a vertical component exceeding the velocity of about 8 m/sec. Drops less than 5 millimetres in diameter are carried upward by the wind, larger drops are unstable and disintegrate. In exceptional circumstances, for example in line squalls and their associated thunderstorms, vertical velocities of this magnitude are undoubtedly reached. If for any reason the exceptional vertical velocity is checked the whole mass of suspended water falls and under such circumstances we get very intense local rainfall. The water, which had circumstances permitted, would have been precipitated gradually and over a wide area falls in a brief interval in a restricted area, often with disastrous consequences to the locality affected. We may note in this connexion that an ascending component of 8 metres per second does not need to persist for very long to carry the air up to the level where the temperature is below the freezing point. The water drops produced by the condensation are then frozen solid and are no longer subject to disintegration but fall as hail. Such intense convection results also in violent electric effects giving rise to thunderstorms.

CHAPTER XII

THE VARIATION OF TEMPERATURE IN THE TROPOSPHERE

AFTER this preliminary consideration of the changes of temperature associated with the vertical motion of air let us examine some of the results obtained from the observations of temperature in the free atmosphere obtained with the aid of balloons or kites. The general features of the variation of temperature with height are represented in Fig. 42, which gives the results of 45 soundings made in the years 1907–8 in the manner which has been briefly described on p. 48. We see at a glance that all ascents agree in showing a fairly rapid lapse rate up to about 10 kilometres while above that level the lapse rate becomes negligible. Confining our attention for the moment to this portion of the atmosphere in which the lapse rate is appreciable and looking more closely into the diagram we find that in the first 2 kilometres from the surface the records of the individual ascents form an almost inextricable tangle, indicative of great complexity of structure. Above the level of 2 kilometres the curves begin to sort themselves out and between the level of 3 and 9 kilometres, though individual ascents still show irregularities, there is as a rule a steady decrease of temperature with height, the lapse rate being ·6 or ·7 degree per hundred metres, rather more than the adiabatic rate for unsaturated air. In the lower part of the diagram, though there is obviously on the average a decrease of temperature with height individual ascents may show almost any features. Some curves in the diagram show a lapse rate

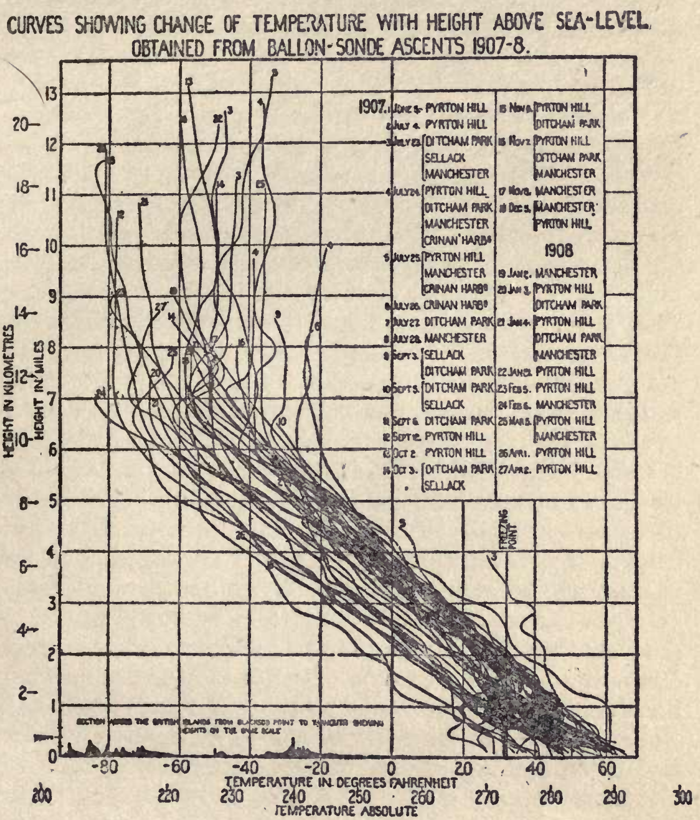


FIG. 42

The separate curves represent the relation between temperature and height in miles or kilometres in the atmosphere. The numbers marking the separate curves indicate the date of ascent at the various stations as shown in the tabular columns. The difference of height at which the stratosphere is reached, and the difference of its temperature for different days or for different localities, is also shown on the diagram by the courses of the lines.

which is practically equal to the adiabatic rate for dry air, others show hardly any change with height, while in many the temperature actually increases with increasing height during a part of the ascent.

We should naturally expect the thermal structure of the lower layers of the atmosphere to be very complex and especially so over the land when we reflect how the surface is warmed and cooled by the alternation of day and night and how the course of that heating and cooling is further complicated by the presence or absence of cloud. We may here draw attention to an important difference between the cases of heating and cooling. If the ground be heated the air in contact with it becomes warmed and therefore less dense so that convection is started. The air rises and its place at the surface is taken by colder air which is warmed in its turn. The heating is thus not confined to the layers in actual contact with the surface but there is a steady upward transference of heat and a tendency to set up a lapse rate which may approach very closely to or even reach the adiabatic rate in a considerable thickness of atmosphere. There is no corresponding upward transference of cold. When the ground cools at night the air in contact with it cools, but it does not ascend in consequence. On the contrary unless carried away in a wind, it remains where it is or if the surface be sloping the coldest air tends to trickle down the slope and collect in the valleys. The cooling of the air in contact with the surface may be very intense but only a comparatively thin layer is affected. On calm nights the temperature recorded by a thermometer exposed an inch or so above the ground may, on occasions, be ten or more degrees lower than that measured in a neighbouring thermometer screen four feet above the ground level. In this way severe ground frosts inflicting much damage on low growing crops may occur at the bottom of valleys while similar crops growing higher up on the hill sides may escape.

This process of drainage of cooled air down sloping ground on a calm cloudless night very frequently leads to the formation of fog, in or near the bottom of valleys or even in hollows in the level surface of the ground too shallow to be referred to as valleys. The cooled air mixes with somewhat warmer air which it meets in the course of its downward drift and may produce a mixture in which the temperature is below the dew point. Condensation in the form of fog results. A certain amount of eddy motion promotes and is even essential to the formation of such fogs. If eddying could be eliminated the moisture condensed from the air which is chilled by contact with the cold ground would be deposited almost entirely as dew or hoar frost, but eddy motion promotes mixing with somewhat higher layers and if the final temperature of the mixture is below the dew point the condensation takes place in the form of fog on dust particles floating in the air. In this way deep and extensive valleys may become entirely filled with fog. The very presence of the fog tends to its own preservation, for much of the solar heat which would otherwise reach the ground and set up convection from below is reflected from the upper surface of the fog screen.

The reversal of the normal lapse rate for temperature above the fog helps to confer special stability on the lower layers so that under suitable general conditions local fogs may persist for days. Convection is impossible until the inverted lapse rate has been gradually destroyed and while it lasts the lower layers are practically cut off, as by a lid, from the rest of the atmosphere. The fog therefore persists and in urban areas may increase steadily in intensity by the further pollution of the air by smoke and dust. In a typical London fog of this class occurring on March 6th, 1902, the minimum temperature recorded on the Victoria Tower of the Houses of Parliament, 400 feet above the ground, was 7·5 degrees F. higher than the minimum measured in a thermometer screen 4 feet above the ground.

Increase of temperature with height, so-called *inversion* in and above a fog is not confined to land fog caused by the cooling of a land surface by radiation. It occurs also in sea fogs formed by the passage of warm air over colder water. Much light has been thrown on the formation of such fogs by G. I. Taylor's theory of eddy motion. Let us consider the case of a current of air in which the lapse rate is the adiabatic, flowing over a cold surface, say a cold sea. The air in contact with the surface becomes chilled, and its temperature falls below that of the air above it, but eddy motion induces mixing of the cold surface air with the warm air above, so that the cooling is not confined to the lowest layers of air, as we might at first sight expect from the absence of convection in the ordinary sense. By a sort of churning process the cooling is gradually extended upwards and the longer the process continues, *i.e.* the longer the "fetch" of the wind, the greater the height at which the influence of the cold surface can be traced. Curves 1 to 5 of Fig. 43 show the distribution of temperature in the vertical calculated by Taylor for a current of air in which the original lapse rate is the adiabatic, passing over cold water for different intervals. The diagonal line represents the original adiabatic state. Curve 1 represents the distribution after 1⅜ hours of flow and curve 2 that after 5½ hours on the assumption that the air velocity is 10 miles per hour, and that the surface air is cooled one degree (absolute) in a run of 30 miles. In the examples shown by curves 3, 4, and 5 the rate of cooling at the surface is taken at 1 degree in 60 miles and the curves show the calculated distribution of temperature after 5½, 22 and 87 hours respectively.

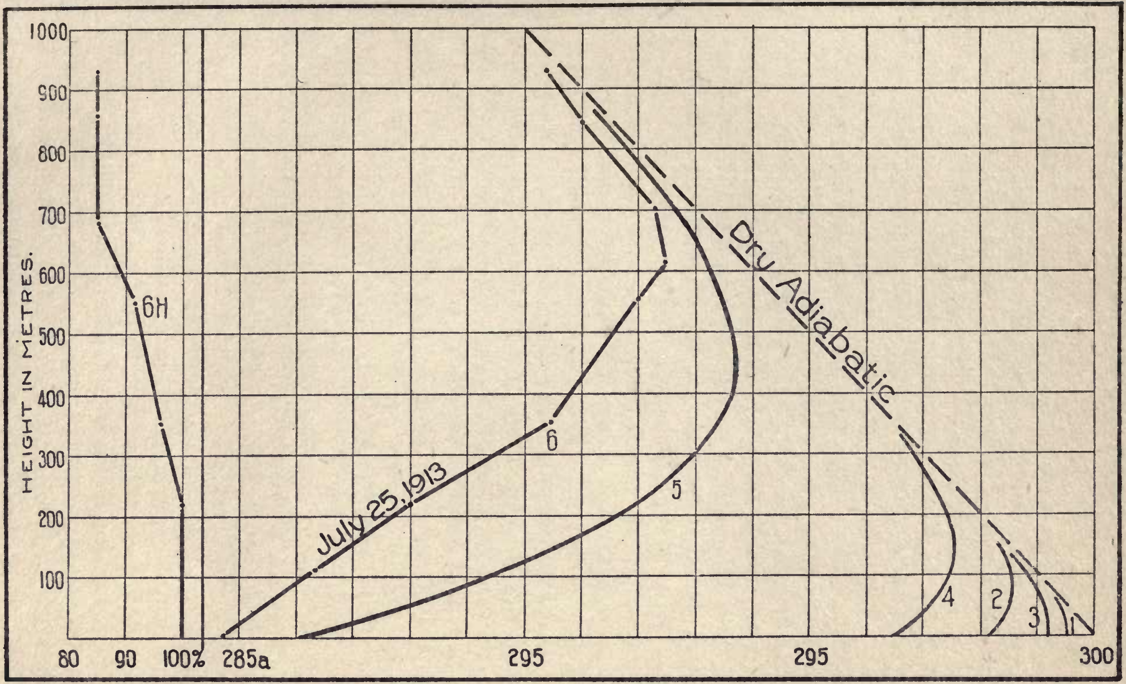
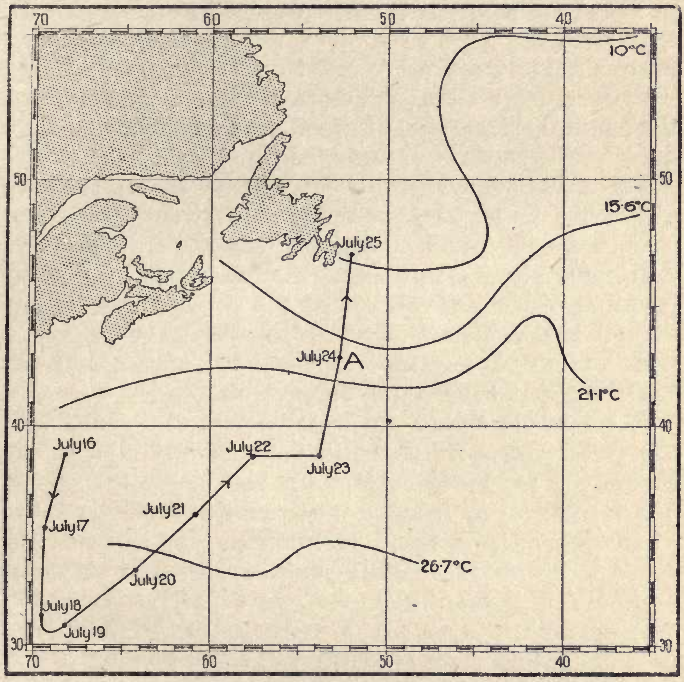


Diagram illustrating Lapse rate when a warm wind blows over a cold surface.

FIG. 43

The general features of these curves are very similar to curve 6 which represents the actual distribution as observed in a kite ascent made during a thick fog from the deck of the s.s. *Scotia* off the Banks of Newfoundland on July 25th, 1913. The diagram may be taken as typical of the conditions prevailing in such fogs. For this particular occasion Taylor was able to trace the previous history of the air by working out its trajectory from the weather maps in the manner described in Chapter VII.

The position of the *Scotia* at 3 p.m. on July 25th, the day of the ascent was eastward of Cape Race. The wind at the time was south by west and had been blowing steadily with a velocity of about 10 miles per hour. We may, therefore, assume that 24 hours earlier the air in which the ascent was actually made was some 240 miles away in a direction south by west, at the point marked A in the diagram (Fig. 44). Making use of all available observations from ships in combination with the published maps of the American Weather Service the probable course of the air during the previous days was traced. It is shown in Fig. 44 the positions for successive days being indicated by the appropriate dates. The chart also shows the temperature of the sea surface, by means of isotherms computed from observations reported during the third week of July by passing steamers. We see from the diagram that from July 16th to 18th the air was moving towards the south towards a region when the temperature of the sea water was high, so that during this part of its course the air current was being warmed from below, a process which would set up in it a lapse rate approaching the adiabatic. From the 18th to the 19th the air remained almost stationary in a region where the temperature of the surface water as observed by a steamer was 81° F., 27° C. or 300 absolute. From July 19th onwards the air was carried into more northern latitudes. At first the motion was directed towards the north-east, and after July 23rd towards north-north-east or north-by-east. The further north it got, the colder the water over which it found itself so that



Trajectories of Air over the Atlantic, —— July, 1913.

FIG. 44

during this part of its flow the air current was being cooled from below under circumstances similar to those for which the lapse curves 1 to 5 of Fig. 43 have been calculated. By 3 p.m. on July 25th the air at the surface had been cooled through about 30° F. to within a degree or so of the temperature of the sea surface observed from the *Scotia.* Curve 6 of Fig. 43 represents the temperature distribution in the vertical as revealed by the kite ascent. We see that the cooling had been propagated upwards by the eddying process to a height of 700 metres or 2,300 feet. Above that level the lapse rate was approximately adiabatic up to the highest level reached by the kite, which was slightly greater than 900 metres. In confirmation of the general correctness of the history of the air current during the ten days under consideration Taylor points out that if the line representing the adiabatic lapse rate above 700 metres be produced as in the dotted line, it cuts the base line of the diagram at a point representing a temperature of 299a, practically the temperature of the sea surface, at the point where the air turned northwards on July 19th. The cooling effect thus extended upwards to the level of 700 metres but the fog did not reach to anything like this height. The curve 6H shows the distribution of humidity in the vertical recorded during the ascent. It shows saturation from the surface up to 210 metres, about 700 feet, and this may be taken as the vertical extent of the fog. Up to this level the churning due to eddy motion has produced an atmosphere containing more moisture than is required to saturate it at the prevailing temperature; the excess of moisture is therefore condensed as fog. Above 210 metres the cooling is not yet sufficient to chill the air below the dew point. The height to which the fog may extend in fogs formed in the manner just described varies considerably. It is frequently so shallow that the mast of a ship is in bright sunshine while the deck is in moderately thick fog; on other occasions the fog layer may extend to 2,000 or 3,000 feet, about 1,000 metres.

Inversion of the normal lapse of temperature is by no means confined to the air immediately above the ground. Layers in which the lapse rate is zero or is even inverted, *i.e.* layers in which there is increase of temperature with increasing elevation, are often found in the free atmosphere. Close scrutiny of the diagram on p. 131 will show many such, especially in the lowest 2 or 3 kilometres, but they occur also higher up. They vary in thickness but are generally several hundred metres thick. They are found almost invariably above clouds of stratus or strato-cumulus type, *i.e.* clouds that are arranged in regular layers or rolls. They are absent above cumulus clouds, isolated cloud patches with rounded tops. Such inversions, by acting as a barrier to the upward penetration of convection cut off the air below them from mixture with the air above. An inversion acts as a sort of lid fixed over the lower layers. Such a lid may sometimes persist for days and the accumulation of smoke and dust stirred up from the surface, give rise to a haze, materially restricting the range of vision. The conception of eddy motion may help us once more in obtaining an insight into the origin of some inversion layers. Let us consider what may happen if a current of air flow over a surface warmer than itself. We assume that the original lapse rate for temperature is less than the adiabatic. Eddy motion over the warm surface would be more vigorous than over the cold fog forming surface considered above in discussing sea fogs. The result of the eddy motion is to induce an approach to the adiabatic lapse rate in the layers affected by it, the upper part of such layer becomes cooled, the lower warmed. The temperature of the upper part of the layer affected by eddy motion may thus become lower than that in the undisturbed air above, in other words there is an inversion. If this process takes place in air containing sufficient moisture the cooling due to the vertical motion associated with the eddying may reduce the temperature below the dew point and in that case cloud will form at the top of the eddying layer. If the distribution of water vapour in the air be uniform to start with such cloud production will commence at a definite level and thus we may get a very definite cloud layer with an inversion above it. The extensive layers of stratus cloud associated with some anti-cyclones perhaps owe their origin to some such process as this. Such clouds sometimes attain considerable density and their appearance is often suggestive of rain, but the inversion above them prevents the more vigorous convection required for appreciable rainfall. Occasionally a persistent drizzle may develop from them.

CHAPTER XIII

THE STRATOSPHERE

HAVING briefly examined the complexities met with near the surface, we turn to the distribution of temperature at great altitudes. We find there materially simpler conditions. All the curves of Fig. 42 show a sudden change in the lapse rate for temperature at a level which varies between 8 and 12 kilometres. Up to that level temperature decreases with height at a rate which is in general somewhat less than the adiabatic lapse rate for dry air. Above this critical level there is no further change of temperature with height or perhaps it would be more correct to say that the changes are negligible by comparison with those shown by the lower layers. The ascents depicted in the diagram are by no means peculiar in this respect. The same general features have been met with in all parts of the world and teach us that as far as temperature is concerned the atmosphere is divided into two parts, a lower shell in which there is fairly rapid lapse of temperature and an upper shell in which there is what for brevity we may call zero lapse rate.

If we consider any one ascent we find that the transition from the régime of falling temperature to that of insignificant change with height is generally, though not invariably, abrupt; it is often marked by a slight bending back of the curve indicating an increase of temperature with height. The level at which the change takes place is not so definite. In the ascents represented in the diagram it varies roughly between the limits of 8 and 12 kilometres. In the majority of ascents the sharp bend occurs near the 10 kilometre level. The existence of this upper layer in which temperature conditions are so very different from what they are nearer the surface was discovered by Teisserenc de Bort in 1899. At first meteorologists were inclined to doubt the reality of the phenomenon and to attribute the recorded cessation of fall of temperature with height to faulty methods of observation, which could be traced to the absorption of solar radiation by the thermograph, but the fact that the same abrupt change was shown by ascents conducted at night, when this source of error was by the nature of the case, eliminated, and that the observations were consistent among themselves soon convinced all that the observations were not at fault and that we have to deal with a genuine change in the régime of temperature at the level of about 10 kilometres.

The upper layer was at first generally referred to as the isothermal layer but that name is not a good one. The upper part of the atmosphere is not a region of uniform temperature as that name would imply. We get a more adequate representation of the distribution in it if we consider our problem in three dimensions and picture to ourselves the isothermal surfaces. In the lower layer, below 10 kilometres, these surfaces are roughly parallel to the surface of the earth. Thus from the diagram, Fig. 42, we may conclude that the isothermal surface of temperature 273a, the freezing point of water, is generally met with over this country at a height less than 2 kilometres, only two of the ascents represented in the diagram show it at a higher level. Above 10 kilometres the isothermal surfaces are no longer approximately horizontal but more nearly vertical.

In order to avoid the erroneous conception associated with the name isothermal layer Teisserenc de Bort suggested the names *stratosphere* for the outer shell of the atmosphere in which there is no material change of temperature with height and *troposphere* for the lower portion in which there is considerable lapse. Both names have now been generally adopted.

The discovery of the existence of the stratosphere was one of the most striking results of the investigation of the upper atmosphere and led to a vigorous campaign to investigate the peculiarities of this new phenomenon. As we may see from our diagram the change from troposphere to stratosphere is generally though not invariably sudden, so that in most ascents we may assign a definite level to the commencement of the stratosphere. This level, for which the name *tropopause* has been suggested, shows considerable variations. Over Europe it varies between about 8 kilometres and 13 kilometres, though both these extremes have been exceeded. Observations are still too few to enable us to give final values for the average height of the tropopause for different regions, but there is good evidence to show that it is dependent to a considerable extent on latitude, being lowest near the poles. For example the ascents of the years 1908–1914 gave an average value for the height of the tropopause over Scotland of 9·8 kilometres, whereas ascents for South-east England gave 10·7 kilometres. Similarly for the years 1904–1909 the average height over Petrograd was 9·6 kilometres but in Northern Italy 11·0 kilometres. The suggestion that the surface of separation between stratosphere and troposphere slopes downward from the equator to the poles received striking confirmation from some ascents carried out by Berson in Equatorial Africa on Lake Victoria in which the stratosphere was not encountered until the height of about 17 kilometres was reached, about 5 kilometres above the usual maximum height for Europe.

The temperature in the stratosphere also shows considerable variations. In Fig. 42 the values lie between 240a and 210a. There is a close connexion between the height of the tropopause and the temperature of the stratosphere which we shall have occasion to refer to again later. We need here only draw attention to the fact that when the tropopause is low, the temperature in the stratosphere is relatively high and *vice versa*. This connexion manifests itself also when we consider the variation of the average temperature in the stratosphere with latitude, for in high latitudes the stratosphere is on the average warmer than it is in low ones. For example for a series of ascents over Petrograd the average temperature at 14 kilometres, well above the tropopause, was found to be 223·5a, whereas the corresponding figure for Pavia in North Italy is 217·7a and for Canada 212·5a. In the ascents on Lake Victoria the extraordinarily low temperature 193a was observed in the stratosphere. We are thus confronted with the fact that, if we wish to find the lowest temperature in Nature accessible to human measurement we should look for it in the stratosphere over the equator. The lowest recorded reading at ground level is 213a, observed on the great ice barrier of the Ross Sea on July 6th, 1911, by Captain Scott's expedition.

We may thus picture to ourselves the stratosphere as an outer shell of the earth's atmosphere, the base of which is at its greatest altitude over the equatorial region and slopes down gradually nearer to the surface as we approach the poles. The average temperature is lowest on the equator where the height is greatest and presumably highest in the neighbourhood of the poles.

Fluctuations in the level of the tropopause and of the temperature in the stratosphere take place from day to day and are connected with the changes of pressure observed at the surface. For example, a series of ascents made in England when grouped according to the pressure at the surface gave the following results for each group of pressures:—

Pressure in mb. ..	> 1033	1033–1027	1027–1020	1020–1013
Height of Tropopause (km.) ..	12·3	11·3	11·2	10·7
Temperature in Stratosphere ..	207	216	216	218
No. of Ascents ..	6	14	37	33

Pressure in mb. ..	1013–1007	1007–1000	1000–993	< 993
Height of Tropopause (km.) ..	10·1	10·2	8·6	8·7
Temperature in Stratosphere ..	218	217	220	223
No. of Ascents ..	25	14	5	16

The ascents made under the conditions of highest pressure, which may be taken as representative of conditions near the centre of an anticyclone, give the highest values for the height of the tropopause and the lowest temperatures in the stratosphere and conversely those made under conditions of low pressures, *i.e.* near the centres of cyclones, give the lowest values for the level of the tropopause and the highest temperatures in the stratosphere.

The discovery of the stratosphere and of the main facts of the variations in its level and temperature has given rise to much speculation as to the cause of this peculiar separation of the atmosphere into two shells. We will briefly examine the conditions which must govern the distribution of temperature when a state of thermal equilibrium is attained. Any portion of the atmosphere or more generally, any portion of matter may be regarded as liable to lose heat and to absorb it from its surroundings. Both processes are continually going on and if a state of constant temperature is maintained, that is equivalent to saying that the gain and loss are equal.

If we consider a thin layer of the atmosphere at any selected level, it must be regarded as radiating heat both upwards towards space and downwards towards the earth. At the same time it is exposed to the heat radiating from other sources, part of which is absorbed and part transmitted. We may distinguish various sources of such radiation. Most obviously there is, by day, the radiation from the sun. Also there is the radiation from the earth which differs from the solar radiation in that it is made up of long heat waves and not of short light waves. In addition there are the radiations from surrounding layers of the atmosphere. If a state of equilibrium is maintained, it may be conceived as a balance between these various items.

The laws according to which gases absorb and radiate heat have been made the subject of experimental investigation in the physical laboratory. Some of the earliest experiments of the kind were made by Tyndall. In these experiments the gas under investigation was led into a tube whose ends were closed by plates of rock-salt, that material being selected on account of its transparency to radiation of long wave length, which is absorbed by glass. A source of heat of definite intensity was placed at one end of the tube, and a thermopile, which could measure the amount of radiation which fell upon it, at the other. By conducting the experiment first with the tube empty and subsequently with the tube filled with various gases, and noting the diminution of the thermopile readings produced by introducing the gas a measure was obtained of the fraction of the radiation absorbed by a layer of gas of specified thickness. For example Tyndall found that water vapour and carbon dioxide show large absorption, but oxygen and nitrogen very little. Subsequent investigators have pushed the matter much further. It has been found that the absorption when it occurs is generally selective, *i.e.* that radiations of certain wave length are absorbed while others are transmitted. It follows that the fraction of the total radiation which is absorbed by the gas, must depend on the nature of the source of heat. If the source does not emit the radiation of the particular wave length to which the gas is partial, there is little or no absorption. The percentage of solar radiation absorbed in a given layer of the atmosphere may be quite different from the percentage of the earth's radiation which the layer can absorb because the latter consists of long heat waves, whereas the former consists of short light waves as well as heat waves.

The theory of the application of such experimental results to the case of the atmosphere has been developed by E. Gold. He has found mathematical expressions for the radiation and absorption in the atmosphere which enable him to reason about what we may call the radiation balance sheet for different levels.

The final temperature attained in any layer is not determined exclusively by the balance between absorption and radiation. We have seen above, Chapter XI., that in an atmosphere in which the lapse rate for temperature exceeds the adiabatic, convection comes into operation. If the radiation balance sheet should set up a sufficiently great lapse rate, convection would occur and play its part, as well as radiation, in deciding the final distribution.

Gold states that the necessary conditions for convection to occur below any specified level are that a thin layer of the atmosphere at that level should lose more energy by radiation than it absorbs, and that the inward radiation at the level should exceed the outward. It is clear that these conditions, if persistent, must eventually give rise to convection for the layer itself would be getting steadily cooler, whereas the layers below it would be warmed by the excess of inward radiation. Ultimately the lapse rate in the lower part must reach and then exceed the adiabatic. He finds that there is a height limit to the level below which the balance is such that convection may result. Below this limit the distribution of temperature is governed largely by convection. The lapse rate approaches the adiabatic, which is one degree per 100 metres in the case of dry air or as shown in Fig. 41 in the case of saturated air. The conditions are practically those which observation has shown to exist in the troposphere. Above the limit, convection in the ordinary sense does not occur, and the distribution of temperature is governed by radiation processes. Making certain assumptions as to the constitution of the atmosphere based on the results of observation Gold calculates that the approximate level of the limit below which convection can occur is 10,500 metres, a value which agrees well with the average height of the tropopause in temperate latitudes, as revealed by observation.

The water vapour in the atmosphere plays a very important part in fixing the limits for convection and the calculated values just quoted involve assumptions as to its distribution based on average conditions. Experiment has shown that water vapour is a much better absorber and radiator of heat than oxygen or nitrogen which are the chief constitutents of the air and its presence exerts a marked influence on the balance of radiation. Its presence in the troposphere increases the loss by radiation from this part of the atmosphere and the actual distribution of temperature as found in the troposphere can only be maintained by the communication of heat to the upper part of the troposphere by convection from below. At the low temperature which prevails in the stratosphere there is practically no water vapour and the loss of heat by radiation, in excess of the gain by absorption, such as occurs in the troposphere, cannot take place. This line of argument may be used to give an explanation of the greater height of the tropopause over the tropics as compared with temperate or polar regions. The high temperatures at the surface which prevail in the tropics give rise to an excess of water vapour there, so that the conditions peculiar to the troposphere extend to a greater altitude than they do in regions where there is less vapour in the air.

CHAPTER XIV

THE THERMAL STRUCTURE OF THE ATMOSPHERE

WE have now considered the distribution of temperature in the very irregular layers near the surface and also in the uppermost layer, the stratosphere. It remains for us to survey briefly the conditions throughout the whole range of height accessible to observation. The first table (p. 151), which is taken from the "Computer's Handbook" gives the mean temperature for each kilometre of height from the ground to 15 kilometres for each month over South-eastern England, as computed by Mr. W. H. Dines. The results have been "smoothed" by means of a sine curve to eliminate chance irregularities which may be expected to disappear when a larger number of observations become available.

Possibly some features which have real significance may have been obliterated by this smoothing but we may accept the figures as giving a general representation of the changes due to change of season. July is the warmest month both at the surface and in the stratosphere and January or February the coldest. The amount of the variation with the season changes little up to 8 kilometres. The mean temperature for the whole year for each level and the amplitude of the seasonal variation, *i.e.* the difference between the mean for the year and the highest or lowest points on the smoothed curve representing the seasonal change are as given in second table on p. 151.

Mean Monthly Temperature in Degrees absolute Smoothed, over S.E. England, for each Kilometre.

Height.	*Jan.*	*Feb.*	*Mar.*	*Apr.*	*May*	*June*	*July*	*Aug.*	*Sept.*	*Oct.*	*Nov.*	*Dec.*
Kilometres.												
15	216	216	217	218	220	221	222	221	220	219	218	217
14	16	16	17	18	20	21	22	21	20	19	18	17
13	16	17	17	18	20	21	21	21	20	19	17	17
12	17	17	17	18	19	20	21	21	20	19	18	17
11	17	17	17	18	19	21	22	22	22	21	20	18
10	19	18	18	20	22	24	25	26	26	25	23	21
9	23	22	23	25	28	30	32	32	32	30	27	25
8	28	28	29	32	35	37	39	39	38	35	32	30
7	35	35	36	39	42	45	47	46	45	42	39	36
6	42	41	42	44	48	51	54	54	53	51	47	43
5	49	48	50	52	56	59	61	61	60	57	53	50
4	56	56	57	59	63	66	68	68	66	64	60	58
3	63	63	64	66	68	71	73	73	72	69	66	64
2	68	68	69	71	74	77	78	78	77	75	72	69
1	72	72	73	76	79	82	84	84	82	80	77	74
Ground	76	76	77	80	84	86	88	88	86	84	80	78

The initial 2 is omitted except in the first line.

Height in Kilometres.	*Mean Temperature for the Year.*	*Amplitude of Seasonal Variation.*
15–20	218·8	3·0
14	18·9	2·8
13	18·7	2·5
12	18·8	2·2
11	19·6	2·9
10	22·2	4·1
9	27·5	5·2
8	33·6	6·0
7	40·7	6·3
6	47·8	6·6
5	54·8	6·6
4	61·7	6·2
3	67·7	5·2
2	73·1	5·4
1	78·0	6·1
Ground	81·8	6·2

At the surface the seasonal variation amounts to 6·2 degrees absolute, it decreases to a minimum value of 5·2° at 3 kilometres and then increases again, reaching its maximum value 6·6° at 5 and 6 kilometres, but up to 9 kilometres the values for the amplitude of seasonal change all lie between the limits 5·2° and 6·6°. In the stratosphere the seasonal variation is distinctly less. The smallest value 2·2° falls to the height of 12 kilometres. Above that level the figures show a slight increase, 3·0 being assigned to the height range 15 to 20 kilometres. The difference between summer and winter conditions of temperature in the stratosphere is thus much less than it is at the ground or throughout the troposphere. At the surface the mean temperature for January is 12·4 degrees below the mean for July, in the stratosphere at 12 kilometres the difference is only 4·4 degrees. There is a corresponding seasonal change in the level of the tropopause, the height at which the stratosphere commences. The mean height of the tropopause calculated from the same ascents, 167 in number, for South-east England is 10·68 kilometres and the amplitude of its seasonal variation in level 0·59 kilometres.

We have already drawn attention (p. 145) to the fact that the variations in the level of the tropopause and in the temperature in the stratosphere are closely connected with the changes of pressure observed near the surface. High pressure at the surface is associated with a high level of the tropopause and low temperature in the stratosphere. In an earlier chapter (p. 33) we discussed how the pressure at any level may be calculated from the pressure at the surface and the distribution of temperature between the surface and the specified level. We will now pursue this subject somewhat further and endeavour to obtain an insight into the average distribution of pressure and temperature at different levels in cyclones and anticyclones, The following table, also taken from W. H. Dines's results as set out in the "Computer's Handbook" gives the average values of pressure and temperature for each kilometre up to the level of 15 kilometres for what we may regard as typical cyclonic and typical anticyclonic conditions:

Height in Kilometres.	*Cyclonic.*		*Anticyclonic.*		*Difference between Anticyclone and Cyclone.*	
	Pressure.	*Temperature.*	*Pressure.*	*Temperature.*	*Pressure.*	*Temperature.*
	Millibars.	°A.	Millibars.	°A.	Millibars.	°A.
15	116	—	123	—	7	—
14	135	224	146	215	11	− 9
13	157	226	171	215	14	−11
12	183	225	201	217	18	− 8
11	212	225	235	221	22	− 4
10	247	225	273	226	26	+ 1
9	288	226	317	233	29	+ 7
8	335	227	366	240	31	+13
7	388	232	422	247	34	+15
6	449	240	483	254	34	+14
5	516	248	552	261	36	+13
4	591	255	628	267	37	+12
3	675	263	713	272	38	+ 9
2	767	269	807	277	40	+ 8
1	870	275	913	279	43	+ 4
0	984	279	1031	282	47	+ 3

The cyclonic values are computed from ascents made on selected occasions when the pressure at the surface was very low the mean value for the selected ascents being 984 millibars. The anticyclonic ones are for selected occasions of conspicuously high pressure, giving a mean value of 1031 mb. at the surface. The figures are very instructive.

At 15 kilometres the pressure over the area which is anticyclonic at the surface is still greater than the pressure over the area where cyclonic conditions prevail at the surface, though the difference has decreased from 47 millibars at the surface to only 7 millibars. We may draw the inference that cyclonic circulation must extend to great heights in the atmosphere. The suggestion found in the writings of older meteorologists that at the level of the upper clouds, say 10 kilometres, the pressure is higher over the centre of a cyclone than in the surrounding area, in other words that conditions at that level are anticyclonic, where they are cyclonic at the surface, is not borne out by the facts as now ascertained. The cyclonic structure with its associated counterclockwise circulation of wind must as a rule extend well into the stratosphere. On the average, the difference of pressure between cyclone and anticyclone does not become zero until a height of about 20 kilometres is reached. Above that level, if the temperature conditions of the stratosphere persist, there would indeed be reversal and the cyclone would become an anticyclone with the highest pressure in the centre. The differences of temperature given in the last column of the table are equally instructive. At the surface the temperature assigned to the anticyclones is 3 degrees higher than that for the cyclones. That is a mere chance depending on the arbitrary selection of the ascents from which the figures given in the table are calculated, but if we look at the figures for various levels in the troposphere we find that the lapse rate is considerably greater in cyclones than anticyclones. At 2 kilometres the mean for the anticyclones is 8 degrees higher than that for the cyclones and at 7 kilometres the difference has increased to no less than 15 degrees. Above this level the difference falls off again and above 8 kilometres the decrease becomes rapid. The temperatures for cyclonic conditions show the stratosphere commencing at about 8 kilometres while those for anticyclonic conditions do not show it until the level of 12 kilometres is reached. The horizontal differences of temperature between cyclone and anticyclone, given in the last column of the table change sign between 10 and 11 kilometres. Above 10 kilometres the anticyclone is distinctly the colder, and remains so at all heights for which observations are available.

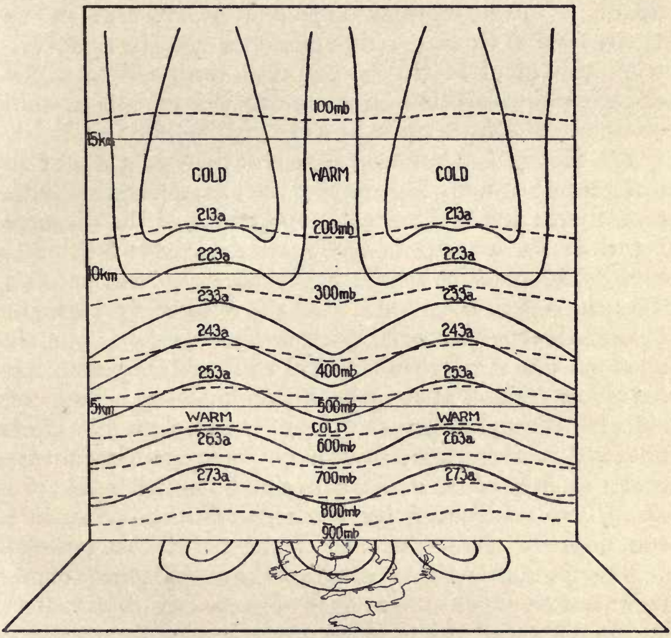
Another theory which at one time was widely held has therefore to be discarded as not being in accordance with the observed facts. A cyclone was, as we now see erroneously, assumed to be a system with a core of warm moisture-laden air and the convection which might be expected to occur under such circumstances was held to be responsible for the maintenance or even for the origin of the system. This view can no longer be held. At levels comprised in the troposphere a cyclone is a region where the air is cold relative to its surroundings. On the other hand an anticyclone, throughout the greater part of its mass is a region of relatively warm air in the troposphere despite the fact that in many anticyclones of winter especially over the continents, the temperature at the surface and in the first few kilometres, may be extremely low. Under such circumstances the intense cold is confined to the layers near the surface which are often cut off from the layers above them by an inversion lid (see p. 139). It is not until the level of the stratosphere is reached that the cyclone becomes the warmer, but the lapse rate then precludes convection as an agent for forming or maintaining the cyclone. The fact that the temperature at moderate altitudes is below the normal in cyclones and above it in anticyclones had been revealed by observations on mountains before observations in the free atmosphere were available but the disturbing influence of the mass of mountain left some doubt as to the applicability of observations on mountains to the atmospheric conditions in the free atmosphere. The newer observations from kites and balloons place the matter beyond doubt.

In order to give a vivid representation of the conditions of pressure and temperature in cyclones and anticyclones we reproduce in Fig. 45 a diagram which has been prepared at the Meteorological Office from a rather more complicated diagram constructed by Mr. W. H. Dines. In this diagram the results of a large number of ascents have been combined to represent a vertical section through the centre of a system consisting of a cyclone with anticyclones on either side of it. The lines shown are the intersection of a vertical plane through the central line of the system with the isobaric and isothermal surfaces. The isobars are shown as broken lines, the isotherms as continuous ones. The distance between the centre of the cyclone where pressure is given as 984 mb. and the centre of the anticyclone where pressure is 1,031 mb. is taken as 500 miles, that is to say, if the cyclone be centred near Holyhead, the anticyclone to the east of it would be centred over North-west Germany. The distribution of pressure and temperature at the centres of the cyclone and the anticyclone respectively are taken from the data given in the table on p. 153. The remainder of the diagram has been constructed by spacing appropriately the results obtained from ascents made under other conditions of pressure at the surface. The words "cold" and "warm" mark the regions where the temperature is respectively below or above the average for the level.

The defect of temperature is greatest at about 7 to 9 kilometres in the cyclonic area. At this level the slope upward from "low" to "high" of the isothermal surfaces, as indicated by the isotherm for 233a, is greatest, but throughout the troposphere the isothermal surfaces suggest horizontal stratification. In the stratosphere on the other hand the isothermal surfaces are approximately vertical as in the case of the isotherm for 223a or inclined to the horizontal at a considerable

DISTRIBUTION OF TEMPERATURE AND PRESSURE IN CYCLONES AND ANTICYCLONES

Mean Values after W.H. Dines.



Solid Curves indicate Temperatures in degrees Absolute [273° being the freezing point of Water]

Broken curves indicate Pressure in Millibars [1000mb being the C.G.S. Atmosphere]

FIG. 45

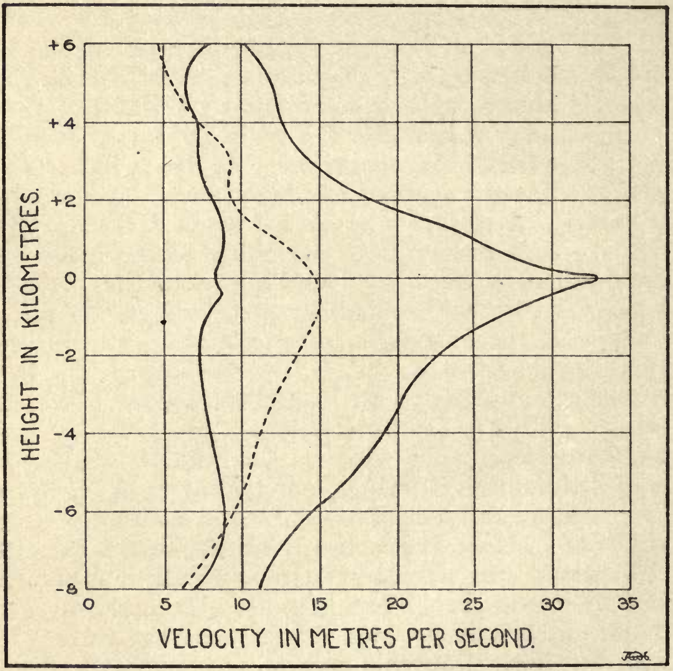
angle as in the case of that for 213a. We may note in passing that the horizontal stratification extends to appreciably higher level over the anticyclone than over the cyclone. For example the isotherms for 223a and

213a are roughly horizontal in the anticyclonic part of the diagram but curve upward rapidly and become what we may call approximately vertical in the cyclonic part.

The diagram suggests that the travel of a depression must be associated with the travel of a dip in the level of the base of the stratosphere which recovers itself again as the depression passes away, and as the temperature in the stratosphere is closely connected with the level of the tropopause the passage of the depression must also be accompanied by considerable changes of temperature in the stratosphere itself.

We may point out also that the vertical scale of the diagram is of course very greatly exaggerated as compared with the horizontal. The width of the diagram, represents a distance of 2,000 miles or about 3,200 kilometres, whereas its height represents only 20 kilometres. To reduce it to a scale drawing the reader must imagine the vertical scale to be compressed to $\frac{1}{160}$ of the size shown. The relative horizontal and vertical dimensions involved should always be borne in mind when considering the structure of cyclones and anticyclones. An analogy is sometimes drawn between such structures and the spin of water in a basin, but such analogy is apt to be misleading for the reason that the ratio of the depth of water to the diameter of the vessel is generally many times greater than the ratio of the height of the atmosphere accessible to our observations to the diameter of a cyclone or anticyclone.

Before leaving the subject of the structure of the atmosphere up to great heights, we will examine the results derived from observations of wind in the stratosphere. Attention has already been called in Chapter X. to the fact that there is very often a marked tendency for the wind to decrease in strength above the level of 10 kilometres, that is to say in the region which we have now learned to recognise as the stratosphere. The number of observations available from heights exceeding 10 kilometres is not very great and moreover, it must be borne in mind that the conditions under which such observations can be made are restricted to the special weather conditions under which pilot balloon ascents reaching great altitudes are possible. In making generalisations we may, therefore, be leaving out of account factors peculiar to cloudy conditions which really play an important part in the scheme of atmospheric circulation. With that proviso, we may summarise the facts by saying that a marked decrease in the velocity of the wind is experienced as the stratosphere is entered. The sudden nature of this change is somewhat masked, if we group the available observations according to height above sea level or ground level and compute the average wind velocity for each kilometre. It is much more obvious if the observations be grouped according to their level above or below the tropopause as has recently been done by G. M. Dobson for the observations recorded in the publications of the International Meteorological Commission on Aeronautics in a paper read before the Royal Meteorological Society. The decrease is most conspicuous in the case of strong winds and indeed is practically absent in the case of light winds. In the case of strong winds, it is shown equally whether the tropopause is high and whether it is low, or whether the pressure at the surface is high or low. Dobson's results are summarised in Fig. 46 which shows the mean variation of the wind velocity with height above or below the tropopause. The available observations are classified into three groups. The first group represented by the curve on the left includes all observations in which the wind velocity in the highest two kilometres of the troposphere was less than 13 metres per second. These occasions of relatively light winds show no change of velocity on passing from troposphere to stratosphere. On occasions of stronger winds the case is different. The dotted curve includes all occasions on which the velocity in the highest two kilometres of the troposphere was between 13 and 19 metres per



Decrease of wind in the Stratosphere.

FIG. 46

second. These show a decided maximum velocity at the level of the tropopause, but the effect is much more conspicuously shown by the third curve which is constructed for the occasions of strong winds at the top of the troposphere on which the velocity in the two kilometres just mentioned exceeds 19 metres per second. The abrupt transition from the condition of velocity increasing with height which prevails throughout the troposphere is here most marked. This abrupt decrease of velocity must be associated with a decrease in the magnitude of the horizontal pressure gradient which we find indicated by the figures giving the difference in pressure at various levels, between anticyclonic and cyclonic conditions shown in the penultimate column of the table on p. 153.

The results as summarised by Dobson show no corresponding change in the direction of the wind on passing from troposphere to stratosphere. The majority of the ascents show no conspicuous change of direction once the disturbed layers near the surface have been traversed up to the greatest heights reached by the balloons and it is quite impossible to identify the level of the tropopause from the results setting out the variation of wind direction with height. Again we must bear in mind that these ascents are restricted to particular weather types; it is quite possible that the generalisation would not hold good if we could include also the occasions which are impossible for observations with pilot balloons on account of cloud, fog or other causes.

CHAPTER XV

THE ORIGIN OF CHANGES OF PRESSURE

THE results which we have just been considering give us an insight into the general conditions of temperature, pressure and wind prevailing at different levels in cyclones and anticyclones and in the earlier chapters we have discussed the circulation of surface winds in such systems, but the questions of their origin or of the causes of the changes which may occur in their configuration have not been discussed. They remain of great difficulty. Some interesting results which, if they do not solve the problem, at any rate suggest in what direction we may look for a solution have been obtained by W. H. Dines by the application of the statistical methods of correlation to the data for the upper air obtained by observation. The application of the method is not confined to meteorological problems; it may be used as a means of tracing relationships between any quantities capable of numerical measurement. Let us suppose that we have a number of simultaneous observations of two quantities. The statistical problem is to obtain some value which may be regarded as a measure of the extent to which the one quality depends on or is influenced by the other. This the statistician proceeds to do by calculating the *correlation coefficient* between the two sets of variables. This coefficient is such that it has a maximum value of unity if the two quantities vary in strict proportion. The nearer the approach of the coefficient to unity the stronger the evidence for a definite connexion between the two variables. For example, if our two sets of variables were daily readings of two neighbouring rain-gauges we should expect a very high coefficient, for the two measurements being compared are obviously closely connected, but if we compare, say, the temperatures in an artificially heated room with those measured out of doors we should find a very low value for the coefficient; indeed if the stoker were successful in maintaining a constant temperature indoors we should expect the coefficient to be near zero. What value of the coefficient may be regarded as significant of direct connexion depends on the number of observations used in the calculation and other circumstances but generally speaking statisticians regard a value exceeding ·5 as evidence of connexion. It does not necessarily follow that the two quantities measured stand to one another in the relation of cause and effect, both may be dependent on some third cause, nor does the coefficient enable us to say which variable should be regarded as cause and which as effect in cases where a causal connexion is established.

As an illustration of the meteorological application of the method we may quote a table of correlation coefficients from the 'Computer's Handbook," which is taken from some results given by Mr. W. H. Dines. The table gives the correlation coefficients between the pressure and temperature for each kilometre up to a height of 13 kilometres calculated from all available ascents of registering balloons made at three stations in Southern England. It is as follows:

CORRELATION COEFFICIENTS BETWEEN PRESSURE AND TEMPERATURE AT HEIGHTS UP TO 13 KILOMETRES.

Height in Kilometres.	0	1	2	3	4	5	6	7	8	9	10	11	12	13
Jan. to March	– ·02	·54	·82	·79	·86	·85	·84	·87	·91	·81	·35	– ·32	– ·38	– ·37
April to June	·14	·28	·49	·79	·89	·89	·92	·87	·81	·45	·20	– ·12	– ·24	– ·01
July to Sept.	– ·02	·31	·56	·72	·75	·81	·83	·87	·87	·88	·43	– ·08	– ·41	– ·19
Oct. to Dec.	·33	·56	·76	·77	·83	·87	·85	·85	·86	·78	·29	– ·24	– ·34	– ·50

At the ground level the correlation coefficients are all exceedingly small, indicating that there is no connexion between our two variables ; any temperature may be associated with any pressure, in other words a knowledge of the reading of the barometer is absolutely no guide in guessing the probable temperature. For the level of 1 kilometre the coefficients are all also small; for the summer months they are well below the critical value ·5, and we may infer that there is no connexion between pressure and temperature. At higher levels conditions are very different. Between the levels of 3 kilometres and 8 kilometres, *i.e.*, throughout the greater part of the troposphere the coefficients are greater than ·7 and in some instances exceed ·9. Here is evidence for a very close connexion, to which attention has already been drawn on p. 114. It means that if we know the pressure at any such level for a particular occasion and find that it is above the average for that level then we may have strong reason for expecting that the temperature at the level will also be above the average and the nearer the approach of the correlation coefficient to unity the greater will be the approach to proportionality of the departures of pressure and temperature from their respective averages.

If we examine the table of coefficients we notice that at 10 kilometres they are quite small again. At that level, which is near the boundary between stratosphere and troposphere but sometimes within the one and sometimes within the other, the connexion breaks down. The fact that the pressure at that level may be known does not help us in the least in estimating the probable temperature. That is quite what we should expect from the considerations set out in Chapter XIV., for we saw there that the temperature in the stratosphere depends very much on the height of the tropopause, and the level 10 kilometres may be sometimes above and sometimes below that critical layer in the atmosphere. At 11, 12 and 13 kilometres we are practically always in the region of the stratosphere. We notice that the sign of the correlation coefficient has become negative, indicating that pressure above the average is now inclined to be associated with temperature below the average, but the values of the coefficient are too small to be significant.

In another paper Mr. Dines has applied this method of analysis to the results of a large number of observations of the upper air, obtained with registering balloons. He has selected the following five variables for calculation:—

(1) The pressure at the surface, P_s.

(2) The mean temperature of the air column between the levels 1 and 9 kilometres, T_m.

(3) The pressure at the level 9 kilometres, P_9.

(4) The height of the tropopause, H_c.

(5) The temperature at the base of the stratosphere, T_c.

The pressure at the level of 9 kilometres was selected as representing the conditions near the top of the troposphere, but generally below the level at which the stratosphere begins.

In addition to calculating the correlation coefficients for each pair of the variables he has calculated for each what statisticians call the *standard deviation*, a quantity which may be regarded as a measure of its variability. He finds that the standard deviations for (1) the pressure at the surface and (3) the pressure at 9 kilometres are 9·4 and 9·2 respectively. These values may be regarded as practically identical from which it follows that the variations of pressure at the level of 9 kilometres are on the average as great as those at the surface. We are, therefore, forced to conclude that the suggestion put forward by the older meteorologists that fluctuations of pressure are greatest at the surface and become less intense as higher altitudes are reached is not borne out by the observations of the lower 9 kilometres. Changes are as great and as complex at 9 kilometres as at the surface.

Turning next to the results obtained by the method of correlation, Mr. Dines gives the mean coefficients calculated for these five variables as follows:

P_s and T_m	·46	T_m and P_9	·92	P_9 and H_c	·83
P_s and P_9	·66	T_m and H_c	·78	P_9 and T_c	−·49
P_s and H_c	·69	T_m and T_c	−·39	H_c and T_c	−·65
P_s and T_c	−·59				

All these values are high and may be taken as indication of a real connexion between the quantities. The negative sign when it occurs indicates that the connexion is an inverse one. Particular interest attaches to the very high value ·92 of the correlation coefficient between T_m the mean temperature of the layer from 1 to 9 kilometres and P_9, the pressure at 9 kilometres. That there should be a connexion between these two is not surprising for P_9 is obtained by a calculation involving T_m but when due allowance is made for this there remains strong evidence that T_m, the mean temperature of the column of air extending from 1 to 9 kilometres is influenced in a direct way by P_9, the pressure at 9 kilometres. Important results follow from this for P_s, the pressure at the surface is made up of the pressure at 9 kilometres *plus* the pressure due to the weight of the column of air between the level of 9 kilometres and the surface. We are thus led on to the conclusion that the pressure at the surface is governed mainly by the pressure at 9 kilometres. If this be so it follows that we must seek the causes giving rise to changes of pressure at the surface at the level of 9 kilometres rather than at or near the surface. The suggestion, which was generally accepted as true until the observations of the upper air collected in the past 20 years taught us otherwise, that the cyclones and anticyclones shown on our maps of surface pressure, are caused by differences of temperature and density in the lower 10 kilometres of the atmosphere likewise falls to the ground.

It seems more probable that the main features of the distribution of pressure are dictated from above and the causes of modifications in the distribution may also have to be sought for at high rather than at low levels.

This suggestion may perhaps help us to understand why the many attempts made in the past to explain the conditions which influence the motion or development of cyclones and anticyclones have been comparatively barren of results. We have perforce had to confine ourselves to a study of surface conditions, whereas the conditions at or above 9 kilometres are really the decisive factors. The conditions of temperature and humidity prevailing at low levels no doubt determine whether a given sequence of changes imposed from above shall give rise to rain, hail or snow, or merely to the development of a certain amount of cloud, but for the cause of larger changes in our pressure maps we must look to higher levels.

CHAPTER XVI

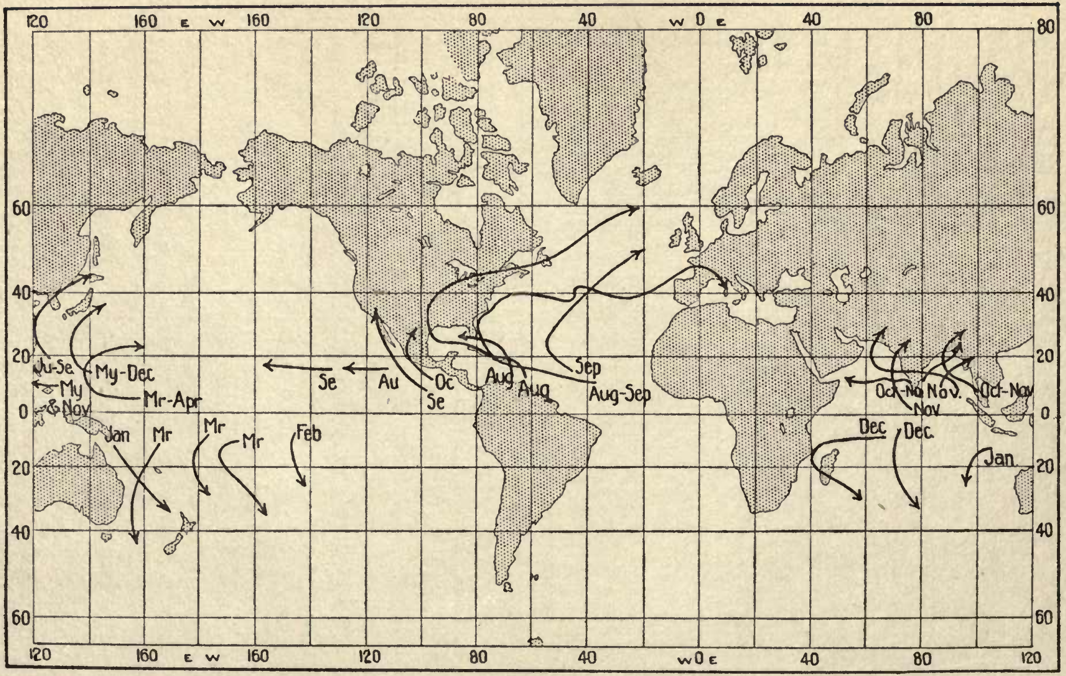
TROPICAL REVOLVING STORMS AND TORNADOES

THE depressions of temperate latitudes which we have had occasion to discuss at some length are generally characterised by a fairly rapid rate of travel. The rate of displacement of the centre of a cyclone is generally of about the same order of magnitude as the velocity of the winds in the system, so that the ratio of the latter to the former rarely exceeds 3. In the case of the depressions of March 24th, 1902, and November 11th, 1901, which we considered in detail in connexion with the trajectories of air in moving cyclones the ratios were 1·1 and 3·0 respectively, and we may note in passing that the latter has been classed as a typical "slow traveller." The trajectories in both these depressions showed long stretches of air motion along paths of but little curvature so that we were justified in many cases in considering only the geostrophic component of the gradient as affecting the conditions of motion. For example the trajectories of the westerly winds on the southern side of the centre of the depression of March 24th show an almost straight flow from west to east.

There are, however, other cases in which the geostrophic component is of relatively slight importance and the air motion is controlled almost exclusively by the cyclostrophic component.

Typical examples of such motion are furnished by the revolving storms of tropical regions. The systems are known by different names in different parts of the world. In the West Indies and among the islands of the South Pacific Ocean such storms go by the name of hurricanes, in the North-west Pacific Ocean and China Sea they are called typhoons and in the Indian Ocean cyclones; but the general features are the same in all regions where they occur. They are rarely if ever, met with actually at the equator, but appear to originate within 5 and 6 degrees of latitude on either side of the line. In these low latitudes their paths are directed from east to west, but there is generally a tendency for drift into higher latitudes and on reaching the 20th parallel or thereabouts the path has been deflected towards the north in the Northern hemisphere and towards the south in the Southern hemisphere, while as higher latitudes are reached the motion becomes directed towards the east. Fig. 47, transcribed from the "Barometer Manual for the Use of Seamen," shows selected tracks for different parts of the world which may be regarded as typical. There is good reason for supposing that some at any rate of these tropical revolving storms when they pass beyond latitude 40° gradually assume the characteristics of the depressions of temperate latitudes. Quite a number of cases are on record, in which West Indian hurricanes have been traced across the Atlantic Ocean to the coasts of the British Isles, which they reached as ordinary cyclonic depressions, often of no great intensity.

While in the tropics the rate of motion of these systems is generally distinctly slow as compared with the rate of advance of depressions in our latitudes. It is generally under 10 miles per hour, though large velocities are occasionally met with. Not infrequently the system remains almost stationary for a day or so. The wind associated with the circulation is often very strong, and wind velocities of over 50 miles per hour must be of common occurrence, and it



Tracks of Tropical Storms

FIG. 47

may safely be said that the ratio of the velocity of the wind to the rate of motion of the storm centre is very much greater than it is in the depressions of our latitudes and in these systems we have a near approach to the gradual indraught of revolving air towards the centre of the system, which the older

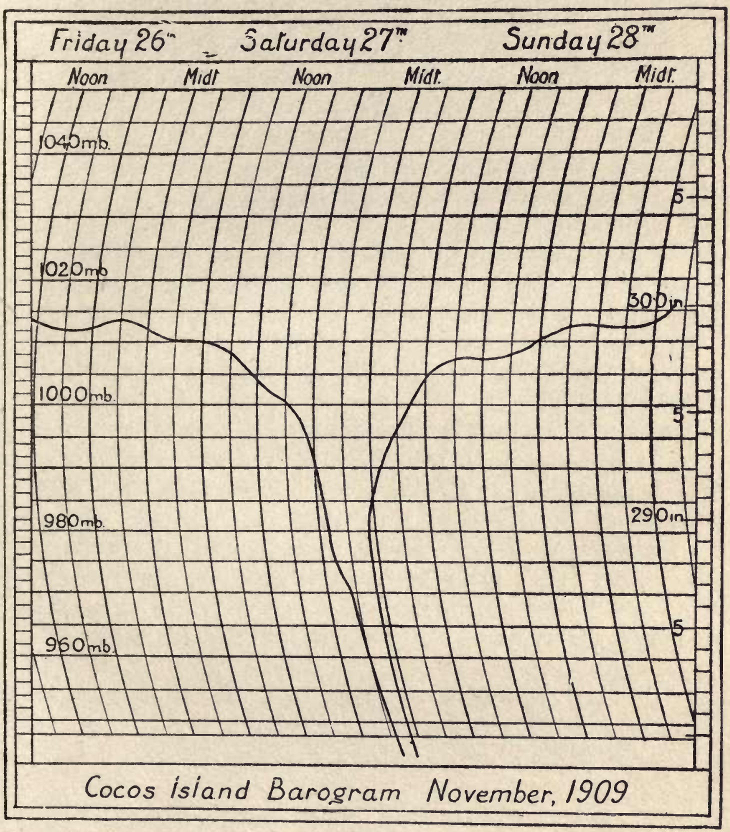


FIG. 48

meteorologists pictured to themselves as the essential feature of a depression. Even in these cases we must be on our guard against forming an analogy between a cyclone and laboratory experiments on vortices or revolving liquids. If we imagine a stationary cyclone, 50 miles in diameter and assume that the winds at a distance of 25 miles from the centre have the high velocity 100 miles per hour, it would take about 1½ hours for the wind to circumnavigate the storm, whereas in a laboratory experiment the time of a revolution is generally measured in seconds rather than in minutes.

In diameter tropical cyclones seem to be as variable as are the depressions of temperate latitudes. The barometer gradient near the centre is often exceedingly steep as is indicated by the exceedingly rapid fall and rise of the barometer as the system approaches and recedes again. Fig. 48 is a reproduction of a barogram from a station in the direct path of a deep cyclone; the actual minimum of pressure was not reached by the instrument, as it was beyond the range for which it was set.

Tropical revolving storms show a marked seasonal variation in all parts of the world where they occur. The following table, taken from the "Barometer Manual for Seamen" gives particulars of their frequency in different months:

Region and Period.	*Jan.*	*Feb.*	*March.*	*April.*	*May.*	*June.*	*July.*	*Aug.*	*Sept.*	*Oct.*	*Nov.*	*Dec.*	*Total.*
West Indies, 300 years	5	7	11	6	5	10	42	96	80	69	17	7	355
South Indian Ocean, 38 years, 1848–1883	71	61	59	50	19	3	2	—	—	5	25	33	328
Bombay, 25 years	1	1	1	5	9	2	4	5	8	12	9	5	62
Bay of Bengal, 139 years ..	2	—	2	9	21	10	3	4	6	31	18	9	115
China Sea, 85 years	5	1	5	5	11	10	22	40	58	35	16	6	214
Arabian Sea, 1877–1903	—	—	—	1	5	6	—	—	—	2	7	—	21
Bay of Bengal, 1877–1903 ..	—	—	—	1	8	4	4	2	6	8	17	6	56
South Pacific, 1789–1891 ..	36	22	35	8	1	—	—	—	2	1	4	16	125

An example of rotational motion in the atmosphere of a different type is furnished by tornadoes and water spouts. In these the diameter of the rotating mass of air is much less than in cyclones being measured by yards rather than miles, but the wind velocities attained may be very great. The most terrible examples occur in the tornadoes of the central states of the American Union. Professor de Courcy Ward gives the following graphic description of these phenomena*:—

"Briefly stated, a tornado is a very intense, progressive whirl, of small diameter, with inflowing winds which increase tremendously in velocity as they near the centre, developing there a counterclockwise vorticular ascensional movement whose violence exceeds that of any other known storm. From the violently agitated main cloud-mass above there usually hangs a writhing funnel-shaped cloud, swinging to and fro, rising and descending—the dreaded sign of the tornado. With a frightful roar, as of '10,000 freight trains,' comes the whirl, out of the dark, angry, often lurid west or south-west, advancing almost always towards the north-east with the speed of a fast train (20 to 40 miles an hour, or more); its wind velocities exceeding 100, 200 and probably sometimes 300 or more miles an hour; its path of destruction usually less than a quarter of a mile wide; its total life a matter of perhaps an hour or so. It is as ephemeral as it is intense. In semi-darkness accompanied or closely followed by heavy rain, usually with lightning and thunder, and perhaps hail, the tornado does its terrible work. Almost in an instant all is over. The hopeless wreck of human buildings, the dead, and the injured, lie on the ground in a wild tangle of confusion. The tornado has passed by.

"Fortunately for man, tornadoes are short-lived, have a very narrow path of destruction, and are by no means equally intense throughout their course. Their writhing funnel cloud, which indicates the region of maximum velocity of the whirling winds, ascends and descends irregularly. Sometimes it may be seen travelling along with its lower end at a considerable distance aloft, like a great balloon. The tornado is then in the clouds, its natural home. Again, under other conditions, the funnel cloud works its way downward to the earth, like a huge gimlet, its base enlarged by the débris drawn in by the inflowing winds; its violence so intense that nothing can resist it. Where the funnel cloud descends, the destruction is greatest; where it rises there are zones of greater safety. The whirl may be so far above the ground that it does no injury whatever. It may descend low enough to tear roofs and chimneys to pieces. It may come down to the ground and leave nothing standing.

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* *Quarterly Journal Royal Meteorological Society*, vol. xliii., p. 317.

"It is one of the most remarkable things about a tornado that even a very short distance—perhaps only a few yards—from the area of complete destruction close to the vortex, even the lightest objects may be wholly undisturbed."

The tornadoes of the United States are most frequent over the great lowlands east and west of the central and upper Mississippi and of the lower Missouri valleys and to a less marked degree over some of the Southern States. They may occur in any month and at any hour of the day or night, but show a preference for the warm months, more particularly April to July and for the warmest part of the day, 3 to 5 p.m. The most favourable meteorological conditions for their occurrence are near but as a rule not actually on the trough of cyclonic depressions crossing the American continent. The warm damp winds, which under such circumstances flow north-east from the Gulf of Mexico appear to favour the development of tornadoes.

Tornado-like phenomena are not unknown in

England and we may conclude that they are liable to occur in almost any part of the world, though they do not as a rule attain to great violence outside the specially marked regions of America.

In England such occurrences are perhaps more frequent than is commonly supposed. Many probably go unrecorded or receive only passing notice in a local newspaper, but in many such descriptions of exceptional wind storms we may recognise the distinctive features of the American tornado, more particularly the narrow, straight lane of winds of destructive force, in which damage is done to trees, greenhouses, or hay ricks. Occasionally the tornado may cross a thickly populated area and then the damage is more severe. As an instance we may refer to the South Wales tornado of October 27th, 1913, which was the cause of considerable loss of life and much destruction of property. The course of the tornado could be traced, with characteristic gaps, along an almost straight line from South Devon to Runcorn in Cheshire. In the Devonshire area we hear of little damage by wind, but reports of torrential rainfall were common. On crossing into Wales typical tornado effects were experienced. A narrow straight lane of destruction varying in width from 50 to 300 yards, extended for a length of 12 miles along the Taff valley. Trees were uprooted and houses and factories wrecked and men killed under the falling débris. Eye-witnesses reported the funnel shaped cloud and roaring noise which always accompany American tornadoes. Similar narrow lanes of severe damage were reported further north in Shropshire and again in Cheshire. In all its essentials the occurrence resembled the American tornado, except perhaps that the winds did not attain the extreme violence commonly experienced in America.

At sea the funnel shaped tornado cloud is known as a water spout. As the point of cloud gradually makes its way downward the sea beneath it becomes agitated testifying to strong wind circulation near it. Ultimately the cloud may descend to the surface and the spout then assumes the appearance of a column of water. It generally lasts from about 10 minutes to half an hour. Frequently the top of the spout is observed to be travelling at a more rapid rate than the base so that the column assumes an oblique or bent form. It ultimately parts at about a third of its height from the base and then quickly disappears.

CHAPTER XVII

THE CHEMICAL COMPOSITION OF THE ATMOSPHERE

IN the preceding chapters we have been able to consider the atmosphere as a gas of uniform composition with the addition of a varying amount of water vapour. We have seen how important a part this water vapour plays in meteorological processes but we have seen also that its presence is confined practically to the troposphere. The temperature conditions of the stratosphere are such that the amount of water present there may be considered as negligible.

In most problems with which the meteorologist has to deal, the chemical composition is a matter of no great importance. We may regard the air as a single gas of specified density and specific heat which obeys the laws of expansion under different conditions of pressure and temperature known by the names of their discoverers Boyle and Charles. It is only in considering problems arising from the radiation or absorption of heat that the chemical composition may become important for, as we have seen, the behaviour of different gases is different in such matters. Nevertheless though the meteorologist is not called upon to give it special consideration in his daily work the chemical composition of the air is a matter of general interest. It has long been known that dry air consists essentially of a mixture of the two elemental gases nitrogen and oxygen in the proportion by volume of four parts of the former to one part of the latter. In addition a small amount of carbon dioxide is always present. The powerful methods of investigation of modern chemistry have shown that small traces of other gases are also present. The most important of these is argon, discovered by Lord Rayleigh in 1895. In an elaborate series of experiments on the density of nitrogen Lord Rayleigh found that specimens of the gas prepared from the atmosphere by removal of all traces of oxygen and carbon dioxide invariably had slightly higher density than specimens prepared from chemical sources. The difference though small was so constant that some "impurity" in the atmospheric samples was suspected and eventually he succeeded in isolating a new gas to which the name of argon was given.

More recently still, thanks mainly to methods of experiment which the extremely low temperatures obtainable with liquid air put at the disposal of the chemist, it has been possible to demonstrate the presence of other gases in the atmosphere and to estimate the proportions in which they occur. The following table taken from the "Computer's Handbook," gives the percentage composition of dry air by volume and by weight, the term dry air being now used in its chemical sense of free from all trace of water:—

	By Volume or Pressure.	*By Weight.*
Nitrogen	78·03	75·48
Oxygen	20·99	23·18
Argon	0·94	1·29
Carbon dioxide	0·03	·045
Hydrogen	0·01(?)	·0007(?)
Neon	·0012	·0008
Helium	·0004	31×10^{-5}
Krypton	$\frac{1}{2} \times 10^{-5}$	$1{\cdot}5 \times 10^{-5}$
Xenon	$\frac{1}{2} \times 10^{-6}$	2×10^{-6}

The chemical constitution of the air at the surface of the earth appears to be constant and we have no reason to suppose that it varies from time to time or from place to place, apart of course from obvious local contamination, by factories, volcanoes or other local causes. Indeed we may go so far as to say that the composition is sensibly the same at all levels accessible to meteorological observation, but at very high levels this is not so.

The gases of the atmosphere do not form a chemical compound, they are merely mixed and each gas may be looked upon as contributing its appropriate fraction to the total pressure of the atmosphere in accordance with Dalton's law of partial pressures. At the surface the fraction of the total pressure which each gas contributes is proportional to its partial pressure.

Now we have seen in Chapter II., that if we know the conditions of temperature we may calculate the pressure at any level from the pressure at the surface. We may perform this calculation for each gas separately and find the total pressure at the high level by adding the individual pressures for the constituent gases separately. It is clear that the pressure due to a heavy gas such as carbon dioxide must decrease relatively much more rapidly than that due to a light gas, such as hydrogen does, and we are led to the result that if the gases of the atmosphere be left to distribute themselves in accordance with their respective densities, the proportion of the different gases present at high levels would differ materially from that at the surface. In the region of the troposphere convection currents would prevent such arrangement according to density and would tend to restore uniform mixing, but in the stratosphere there is no convection and we may suppose that the gases may actually distribute themselves according to their density.

The following table, taken from a paper by Humphreys, shows how the composition of the atmosphere at very high levels may differ from that at the surface. It is computed on the assumption that at the surface the air contains 1·20 per cent. of water vapour; hence the figures for the percentage distribution at the surface differ slightly from those given above.

Height in Kilometres.	*Gases.*							*Total Pressure in Millimetres.*
	Argon.	*Nitrogen.*	*Water Vapour.*	*Oxygen.*	*Carbon Dioxide.*	*Hydrogen.*	*Helium.*	
150	—	—	—	—	—	99·73	0·27	·0043
140	—	—	—	—	—	99·70	0·30	·0048
130	—	0·02	—	—	—	99·64	0·34	·0054
120	—	0·10	—	—	—	99·52	0·38	·0060
110	—	0·40	—	0·02	—	99·16	0·42	·0067
100	—	1·63	—	0·07	—	97·84	0·46	·0076
90	—	6·57	—	0·32	—	92·62	0·49	·0090
80	—	22·70	—	1·38	—	75·47	0·45	·0123
70	0·02	53·73	—	4·05	—	41·95	0·27	·0248
60	0·04	78·16	—	7·32	—	14·33	0·15	·081
50	0·08	86·16	—	10·01	—	3·72	0·03	·466
40	0·16	86·51	—	12·45	—	0·88	—	1·65
30	0·22	84·48	—	15·10	—	0·20	—	8·04
20	0·55	81·34	—	18·05	0·01	0·05	—	39·6
15	0·74	79·56	—	19·66	0·02	0·02	—	88·2
11	0·94	78·02	0·01	20·99	0·03	0·01	—	168·
5	0·94	77·89	0·18	20·95	0·03	0·01	—	405·
0	0·93	77·08	1·20	20·75	0·03	0·01	—	760·

It will be noticed that at the level 150 kilometres, where the total pressure is only ·0043 millimetre or about ·0057 millibar, the atmosphere consists of over 99 per cent. of hydrogen. But even at less extreme levels there is appreciable change from the conditions at the surface. Water vapour is, as we have already seen, confined to the troposphere. The carbon dioxide decreases to ·01 per cent. at 20 kilometres, a level which is within the limits reached by balloon records. Argon decreases to ·02 per cent. at 70 kilometres, at which level the proportion of nitrogen has decreased to 53·73 per cent., that of oxygen to only 4·05 per cent. while hydrogen makes up 41·95 per cent., approaching one half, of the atmosphere.

While the final proofs were passing through the press Professor S. Chapman and Mr. E. A. Milne read a most interesting paper before the Royal Meteorological Society (*Quarterly Journal*, vol. xlvi.) in which they brought forward grounds for doubting the existence at great heights of an atmosphere consisting mainly of hydrogen. The amount of hydrogen present near the surface is exceedingly small, and, moreover, the determinations of its amount made by different experimenters differ considerably. These discrepancies may, perhaps, be due to actual differences in the amount present from time to time, such as we might expect if the hydrogen were derived from organic decomposition, volcanoes, etc. They would be accentuated by the opportunities which exist for the removal of free hydrogen by combination with oxygen or ozone. Very possibly these opportunities may suffice for the complete removal of all hydrogen before it diffuses to the stratosphere. Moreover, the spectrum of aurora (height about 100 to 130 kilometres) shows no lines due to hydrogen, and this evidence, though not conclusive, supports the view that hydrogen is absent. If these considerations hold good the composition of the atmosphere above 100 kilometres must be quite different from that set out in the table just given. Helium would be the principal constituent. Chapman and Milne give the following figures:

Height. Kilometres.	*Helium. Per Cent.*	*Nitrogen. Per Cent.*	*Oxygen. Per Cent.*
100	11·4	84·4	4·2
110	31·3	66·0	2·7
120	61·6	37·2	1·2
130	84·8	14·8	0·4
140	95·1	4·8	0·1
150	98·6	1·4	0·0

PRESSURE

Equivalents in Millibars of Inches of Mercury at 32° F. and Latitude 45°.

Inches of Mercury.	*Hundredths of Inches.*									
	0	1	2	3	4	5	6	7	8	9
	Millibars.									
27·0	914·3	914·6	915·0	915·3	915·7	916·0	916·3	916·7	917·0	917·4
27·1	917·7	918·0	918·4	918·7	919·0	919·4	919·7	920·1	920·4	920·7
27·2	921·1	921·4	921·8	922·1	922·4	922·8	923·1	923·4	923·8	924·1
27·3	924·5	924·8	925·1	925·5	925·8	926·2	926·1	926·8	927·2	927·5
27·4	927·9	928·2	928·5	928·9	929·2	929·5	929·9	930·2	930·6	930·9
27·5	931·2	931·6	931·9	932·3	932·6	932·9	933·3	933·6	933·9	934·3
27·6	934·6	935·0	935·3	935·6	936·0	936·3	936·7	937·0	937·3	937·7
27·7	938·0	938·3	938·7	939·0	939·4	939·7	940·0	940·4	940·7	941·1
27·8	941·4	941·7	942·1	942·4	942·8	943·1	943·4	943·8	944·1	944·4
27·9	944·8	945·1	945·5	945·8	946·1	946·5	946·8	947·2	947·5	947·8
28·0	948·2	948·5	948·8	949·2	949·5	949·9	950·2	950·5	950·9	951·2
28·1	951·6	951·9	952·2	952·6	952·9	953·2	953·6	953·9	954·3	954·6
28·2	954·9	955·3	955·6	956·0	956·3	956·6	957·0	957·3	957·7	958·0
28·3	958·3	958·7	959·0	959·3	959·7	960·0	960·3	960·7	961·0	961·4
28·4	961·7	962·1	962·4	962·7	963·1	963·4	963·7	964·1	964·4	964·8
28·5	965·1	965·4	965·8	966·1	966·5	966·8	967·1	967·5	967·8	968·1
28·6	968·5	968·8	969·2	969·5	969·8	970·2	970·5	970·9	971·2	971·5
28·7	971·9	972·2	972·6	972·9	973·2	973·6	973·9	974·2	974·6	974·9
28·8	975·3	975·6	975·9	976·3	976·6	976·9	977·3	977·6	978·0	978·3
28·9	978·6	979·0	979·3	979·7	980·0	980·3	980·7	981·0	981·4	981·7
29·0	982·0	982·4	982·7	983·0	983·4	983·7	984·1	984·4	984·7	985·1
29·1	985·4	985·8	986·1	986·4	986·8	987·1	987·5	987·8	988·1	988·5
29·2	988·8	989·1	989·5	989·8	990·2	990·5	990·8	991·2	991·5	991·9
29·3	992·2	992·5	992·9	993·2	993·5	993·9	994·4	994·6	994·9	995·2
29·4	995·6	995·9	996·3	996·6	996·9	997·3	997·6	997·9	998·3	998·6

PRESSURE—*Continued.*

Inches of Mercury.	*Hundredths of Inches.*									
	0	1	2	3	4	5	6	7	8	9
	Millibars.									
29·5	999·0	999·3	999·6	1000·0	1000·3	1000·7	1001·0	1001·3	1001·7	1002·0
29·6	1002·4	1002·7	1003·0	1003·4	1003·7	1004·0	1004·4	1004·7	1005·1	1005·4
29·7	1005·7	1006·1	1006·4	1006·8	1007·1	1007·4	1007·8	1008·1	1008·4	1008·8
29·8	1009·1	1009·5	1009·8	1010·1	1010·5	1010·8	1011·2	1011·5	1011·8	1012·2
29·9	1012·5	1012·8	1013·2	1013·5	1013·9	1014·2	1014·5	1014·9	1015·2	1015·6
30·0	1015·9	1016·2	1016·6	1016·9	1017·3	1017·6	1017·9	1018·3	1018·6	1018·9
30·1	1019·3	1019·6	1020·0	1020·3	1020·6	1021·0	1021·3	1021·7	1022·0	1022·3
30·2	1022·7	1023·0	1023·3	1023·7	1024·0	1024·4	1024·7	1025·0	1025·4	1025·7
30·3	1026·1	1026·4	1026·7	1027·1	1027·4	1027·7	1028·1	1028·4	1028·8	1029·1
30·4	1029·4	1029·8	1030·1	1030·5	1030·8	1031·1	1031·5	1031·8	1032·2	1032·5
30·5	1032·8	1033·2	1033·5	1033·8	1034·2	1034·5	1034·9	1035·2	1035·5	1035·9
30·6	1036·2	1036·6	1036·9	1037·2	1037·6	1037·9	1038·2	1038·6	1038·9	1039·3
30·7	1039·6	1039·9	1040·3	1040·6	1041·0	1041·3	1041·6	1042·0	1042·3	1042·6
30·8	1043·0	1043·3	1043·7	1044·0	1044·3	1044·7	1045·0	1045·4	1045·7	1046·0
30·9	1046·4	1046·7	1047·1	1047·4	1047·7	1048·1	1048·4	1048·7	1049·1	1049·4

INDEX

Absolute Temperature, 49
Adiabatic, 121
Anemometry, 38
Aneroid Barometer, 26
— —, Errors of, 34
Anticyclone, 5, 156
—, Temperature and Pressure in, 153-157
Aspiration Thermometer, 46
Average Pressure Distribution and Wind Circulation, 67

Backing Wind, 14
Barometer, 24
— Corrections, 25
Beaufort Scale of Wind Force, 37
— Weather Notation, 3
Berson, 143
Bjerknes, 97
Blood Rain, 88
Buys Ballot's Law, 5, 61

Chemical Composition of Air, 177, 180
Chinook Wind, 82
Clearing Shower, 19, 96
Cloud, Anticyclonic, 7, 140
—, Formation of, 123
— Forms, 53
—, Measurement of Height, 52
Computed Wind at 1,500 feet, 109
Convection, 122, 147
Cooling by Eddy Motion, 134
— of Air by Expansion, 81, 122
Correlation Coefficient, 162
— of Pressure and Temperature in Troposphere, 163
Cyclones, 8, 156, 169
—, Pressure and Temperature in, 153-157
Cyclostrophic Wind, 60, 62, 168

Daily Weather Report, 3, 43
Depression, 8
Dines, J. S., 106
—, W. H., 48, 150, 153, 156, 162
Diurnal Variation of Wind, 103, 104
Dobson, G. M., 107, 159

Easterly Type, 64
East Winds, Decrease of Velocity in, 112
Eddy Motion, 102
— — and Fog Formation, 134
Eiffel Tower, Winds on, 106

Faults in Isobars, 94
Foehn Wind, 82
Fog in Valleys, 133
— at Sea, 134
Frequency Distribution of Pressure over British Isles, 29

Gases, Absorption in, 147
Geostrophic Wind, 60, 61
Gold, E., 147
Gradient, 58, 122
— of Pressure, Change with Altitude, 119
— Wind, 60, 63, 108
Gravity, Barometer Correction for, 25
Gustiness, 39

Hail, 129
Height, Determination of, from Pressure, 33
Hurricane, 37, 169
Hydrogen Atmosphere, 180

Increase of Wind with Height, 102, 107, 111

Incurvature, 108
International Weather Symbols, 4
Inversion, 134, 139
Isobars, 4
Isothermal Layer, 142

Lapse, 122
— of Temperature in the Free Atmosphere, 130
Latitude, Barometer Correction for, 25
—, Variation of Stratosphere with, 144
Lines of Convergence and Divergence, 98
Line Squall, 91
Lowest Temperature on Record, 144

Mercury Barometer, 24
Millibar, 28
Mixing, Formation of Fog by, 133
— and Rainfall, 127

Night Frost, 132

Orographic Rain, 82

Paths of Depressions, 13, 16
Pilot Balloon Observations, 41, 109
Pressure Changes, Causes of, 166
—, Distribution at 4,000 metres, 75
— — in January and July, 68, 71
—, Gradient at High Levels, Evaluation from Pilot Balloon Observations, 113, 114
—, Normal and Extreme Variations, 29
—, Reduction to Sea Level, 30
— Tube Anemometer, 38

Radiation, Balance of, 146
Raindrops, Size of, 128
Rain, Formation of, 81,121, 126
Reduction to Sea Level, 30
Registering Balloons, 47
Robinson Cup Anemometer, 38

Scotia, Kite Ascents on board, 136
Scott Expedition, Lowest Temperature, 144
Sea Fog, 134
Seasonal Variation of Temperature at Various Levels, 151
— — of Tropical Storms, 172
Secondary, 19
Spiral Motion of Air in Depressions, 76
Squall Line, 98
Standard Deviation, 165
Steering of Air by Pressure Gradient, 65
Steering Line, 98
Stevenson Screen, 45
Stratosphere, 143
—, Variation of Level and Temperature, 145
—, Wind in, 159
Stratus Cloud, Formation by Eddy Motion, 139
Symbols, International Weather, 4

Taylor, G. I., 102, 108, 134
Teisserene de Bort, 75, 142, 143
Temperature at Great Heights, 141
—, Averages up to 15 Kilometres, 151
—, Lowest Recorded, 144
—, Scales of, 49
Thermal Structure near Surface, 132
Thermometer Exposure, 45
— Scales, 49
Tornadoes, 173
Torricelli, 24
Tracks of Depressions, 16
Trajectories of Air, 78
Tropical Revolving Storms, 169
Tropopause, 143, 152
Troposphere, 143
Trough, 14, 16
Types, Easterly and Westerly, 64, 66
Typhoons, 169

Upper Wind, Measurement of, 41, 43
— Air Supplement to Daily Weather Report, 112
Upward Transfer of Heat by Convection, 132

Veering Wind, 14
V-shaped Depression, 16, 97

Ward, de Courcy, 173
Warming of Air by Descent, 82, 122
Waterspout, 175
Water Vapour and Height of Tropopause, 148
Wedge, 21
Westerly Type, 66
Wind, Estimation of, 36
—, Increase with Height, 102, 107, 111
— in Stratosphere, 159
— Velocity over Grass-land, 101

Zeppelins, Loss of, 8, 13, 115

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